Structural inheritance in the North Atlantic

Christian Schiffer[1,2], Anthony G. Doré[3], Gillian R. Foulger[2], Dieter Franke[4], Laurent Geoffroy[5], Laurent Gernigon[6], Bob Holdsworth[2], Nick Kusznir[7], Erik Lundin[8], Ken McCaffrey[2], Alex Peace[9,10], Kenni D. Petersen[11], Thomas Phillips[2], Randell Stephenson[12], Martyn S. Stoker[13], Kim Welford[9]

[1] Department of Earth Sciences, Uppsala University, Villavägen 16, 75236 Uppsala, Sweden

[2] Department of Earth Sciences, Durham University, Science Laboratories, South Rd. DH1 3LE, UK

[3] Energy & Geoscience Institute (EGI, University of Utah), London W5 2SE.

[4] Bundesanstalt für Geowissenschaften und Rohstoffe (Federal Institute for Geosciences and Natural Resources), Germany

[5] Université de Bretagne Occidentale, Brest, 29238 Brest, CNRS, UMR 6538, Laboratoire Domaines Océaniques, 29280 Plouzané, France

[6] Norges Geologiske Undersøkelse (NGU), Geological Survey of Norway, Leiv Erikssons vei 39, N-7491 Trondheim, Norway

[7] University of Liverpool, School of Environmental Sciences, Liverpool L69 3GP, United Kingdom;

[8] Equinor, Research Centre, Arkitekt Ebbels vei 10, 7053 Trondheim, Norway

[9] Department of Earth Sciences, Memorial University of Newfoundland, St. Johns, Newfoundland, Canada, A1B 3X5

[10] now at: School of Geography and Earth Sciences, McMaster University, Hamilton, Ontario, Canada, L8S 4L8

[11] Department of Geoscience, Aarhus University, Høegh-Guldbergs Gade 2, DK-8000 Aarhus C., Denmark

[12] School of Geosciences, University of Aberdeen, King's College, Aberdeen AB24 3UE, UK

[13] Australian School of Petroleum, University of Adelaide, Adelaide, SA 5005, Australia

Abstract

The North Atlantic, extending from the Charlie Gibbs Fracture Zone to the north Norway-Greenland-Svalbard margins, is regarded as both a classic case of structural inheritance and an exemplar for the Wilson-cycle concept. This paper examines different aspects of structural inheritance in the Circum-North Atlantic region: 1) as a function of rejuvenation from lithospheric to crustal scales, and 2) in terms of sequential rifting and opening of the ocean and its margins, including a series of failed rift systems. We summarise and evaluate the role of fundamental lithospheric structures such as mantle fabric and composition, lower crustal inhomogeneities, orogenic belts, and major strike-slip faults during breakup. We relate these to the development and shaping of the NE Atlantic rifted margins, localisation of magmatism, and microcontinent release. We show that, although inheritance is common on multiple scales, the Wilson Cycle is at best an imperfect model for the Circum-North Atlantic region. Observations from the NE Atlantic suggest depth dependency in inheritance (surface, crust, mantle) with selective rejuvenation depending on time-scales, stress field orientations and thermal regime. Specifically, post-Caledonian reactivation to form the North Atlantic rift systems essentially followed pre-existing orogenic crustal structures, while eventual breakup reflected a change in stress field and exploitation of a deeper-seated, lithospheric-scale shear fabrics. We infer that, although collapse of an orogenic belt and eventual transition to a new ocean does occur, it is by no means inevitable.

Keywords: North Atlantic; Wilson Cycle; plate tectonics; structural inheritance; reactivation; rifting; continental breakup; magmatism; lithosphere;

Contents

List of abbreviations

CNAR – Circum-North Atlantic Region

COB – Continent Ocean Boundary

COT – Continent Ocean Transition

GGF – Great Glen Fault

GIFR – Greenland-Iceland-Faroe Ridge

GIR – Greenland-Iceland Ridge

HBF – Highland Boundary Fault

HFZ – Hardangerfjord Fault Zone

HVLC/HVLCB – High velocity lower crust/ High velocity lower crustal body

IFR – Iceland-Faroe Ridge

JMMC – Jan Mayen Microplate Complex

Moho – Mohorovičić Discontinuity/Crust-Mantle boundary

MTTC – Møre-Trøndelag Fault Complex

NAIP – North Atlantic Igneous Province

SDR – Seaward Dipping Reflector

TZ – Tornquist Zone

TIB – Trans-Scandinavian Igneous Belt

WBF – Walls Boundary Fault

1 Introduction

Structural inheritance has been invoked as an important influence on plate-tectonic processes including rifting, and rifted-margin end-member style (i.e., magma-rich or magma-poor) (e.g. Vauchez et al. 1997; Bowling and Harry 2001; Manatschal et al. 2015; Chenin et al. 2015; Schiffer et al. 2015b; Svartman Dias et al. 2015; Petersen and Schiffer 2016; Duretz et al. 2016), the formation of oceanic fracture zones, transform faults, and transform margins (Bellahsen et al. 2006; Gerya 2012; Doré et al. 2015; Peace et al. 2018b) , magmatism (Hansen et al. 2009; Whalen et al. 2015), and intraplate deformation (Stephenson et al. this volume; Sutherland et al. 2000; Gorczyk and Vogt 2015; Audet et al. 2016; Heron et al. 2016; Tarayoun et al. 2018; Heron 2018).

The inspiration for major concepts of large-scale structural inheritance, such as the "Wilson Cycle" lies in the Circum-North Atlantic region (CNAR) (Wilson 1966; see review by Wilson et al. 2019), where at least two oceans have opened and closed along broadly similar trends (Cawood et al. 2007; Bingen et al. 2008a; Li et al. 2008; Lorenz et al. 2012; Thomas 2018).

The CNAR comprises the North Atlantic Ocean, Labrador Sea-Baffin Bay, Iceland and the surrounding continental landmasses, including Greenland, Scandinavia, the British Isles and northeastern Canada. The lithosphere comprises stable Precambrian continental cores in the interior of Greenland, North America and Scandinavia, while the geology along the continental margins and northern Europe was mainly reshaped in the Phanerozoic (e.g. Peace et al. this volume; Cocks and Torsvik 2006, 2011; St-Onge et al. 2009). The continental margins host a number of failed rift systems, such as the North Sea, the Rockall-Hatton Basins and the Møre-Vøring Basins (Figure 1) (Péron-Pinvidic and Manatschal 2010; Peace et al. 2019). In detail, continental breakup did not always follow earlier rift systems or known orogenic structures and has in some cases broken through seemingly undisturbed cratonic lithosphere. Several aspects of North Atlantic geology remain enigmatic, such as the nature and significance of the North Atlantic Igneous Province (NAIP) and the Greenland-Iceland-Faroes Ridge (GIFR) (Vink 1984; White and McKenzie 1989; Foulger and Anderson 2005; Meyer et al. 2007; Foulger et al. 2019), the development of a spectrum of rifted continental margins (Geoffroy 2005; Franke 2013; Clerc et al. 2018), and the development of the Jan Mayen Microplate Complex (JMMC) (Foulger et al. 2003; Gaina et al. 2009; Gernigon et al. 2015; Blischke et al. 2017, 2019; Schiffer et al. 2018).

Whether rifting, continental breakup and associated magmatism were related to deep, active mantle upwelling (White and McKenzie 1989; Hill 1991) or plate tectonic processes (Nielsen et al. 2007; Ellis and Stoker 2014; Foulger et al. 2019) (the bottom-up and top-down views) is still under debate (Peace et al. this volume; van Wijk et al. 2001; Foulger et al. 2005, 2019; Lundin and Doré 2005). Despite the often-proposed deep, active, buoyant upwellings beneath the CNAR, factors like the thermal state and composition of the crust and mantle, small-scale convection, upwelling, volatile content, and general, pre-existing (inherited) lithospheric and crustal structure may play major roles in the magmatic and tectonic evolution (e.g. King and Anderson 1998; Asimow and Langmuir 2003; Korenaga 2004; Foulger et al. 2005, 2019; Meyer et al. 2007; Simon et al. 2009; Hole and Millett 2016; Petersen et al. 2018; Hole and Natland 2019).

In this contribution, we aim at defining the most important concepts of structural inheritance and review how they may have influenced the structural evolution of the CNAR as a whole. We then take five segments of the CNAR that differ markedly in structural style as examples (Fig. 1) and describe and discuss these further: namely, the Norwegian-Greenland Sea (*segment 1*), where early rifting followed Caledonian crustal trends, but breakup occurred obliquely, is in contrast to the SE Greenland-Rockall-Hatton margins (*segment 2*), where rifting and breakup occurred through seemingly undisturbed cratonic lithosphere, but parallel to the Caledonian trends in the British and Irish Isles, some 500 km to the east. The enigmatic GIFR, a large physiographic high crossing the North Atlantic (*segment 3*) forms a buffer between *segments 1* and *2*. The North Sea (*segment 4*) forms a major failed intracontinental rift system influenced by Variscan, Caledonian and Precambrian inheritance, but never developed into a new ocean. Lastly, in the Labrador Sea and Baffin Bay rifting broke through cratonic lithosphere but seafloor-spreading was abandoned after ~30 Ma (*segment 5*).

2 The Wilson Cycle and the North Atlantic

Tuzo Wilson's famous question of 1966, "Did the Atlantic close and then re-open?" gave rise to the "Wilson Cycle" concept (Wilson 1966; Dewey and Spall 1975; see review by Wilson et al. 2019). In its simplest form, this hypothesis envisages closure and reopening of oceans along former orogens that represent the weakest zones in a disintegrating continent. Applied stresses exploit inherited weaknesses during later rifting events, rather than breaking up continents through their stronger, stable interiors.

This paradigm works well in the Central Atlantic, where the new ocean closely tracks the parallel Appalachians (Thomas 2018). Similarly, breakup between Scandinavia and Greenland generally follows the Caledonian Orogen, but farther south the Iapetus suture is preserved and runs through northern England and central Ireland (McKerrow and Soper 1989; Soper et al. 1992). The rifting thus left significant pieces of Laurentian cratonic crust on Europe's northwestern seaboard including the Rockall-Hatton margin. The Labrador Sea and Baffin Bay cut through pre-existing cratons (the Archaean North Atlantic and Rae cratons) and almost orthogonally across Precambrian orogenic belts (Buchan et al. 2000; Bowling and Harry 2001; St-Onge et al. 2009; Peace et al. 2018b).

It is becoming increasingly clear that the age of inherited structures prone to rejuvenation extends much further back in time than simply the most recent Wilson Cycle. Accordingly, Archaean-to-Palaeoproterozoic structures also guided fragmentation and segmentation of onshore and offshore areas during rifting and continental breakup in the NE Atlantic (Gabrielsen et al. 2018; Rotevatn et al. 2018; Schiffer et al. 2018) and Labrador Sea-Baffin Bay (Peace et al. 2018a; Heron et al. 2019). Recent attempts to formally extend the Wilson Cycle concept have been made, for example by including reactivation of long-lived intraplate inheritance (Heron et al. 2016), by systemising the role of mantle plumes in the Wilson Cycle (Heron, 2018), or by adding systematic "short-cuts" through the Wilson Cycles, such as the closure of failed rift basins (Chenin et al. 2018).

Current understanding of the precise mechanisms that govern rifting and breakup is hindered by ambiguous observations, interpretations, concepts and definitions. The exact location and definition of the continent-ocean "boundary" is often not known due to the presence of magmatic or sedimentary cover and, in many cases, continental margins have wide transition zones (Eagles et al. 2015). High velocity lower crust (HVLC, see Foulger et al., this volume, and Gernigon et al., this volume for discussion) underlying continental margins can have different pre-, syn-, and post-rift/breakup origins, knowledge of which is crucial to understanding thinning, magmatism and the role of structural inheritance during rifting. Local mantle upwellings associated with small-scale convection or diapirism and magmatic intrusions prior and during continental extension and breakup may have a crucial role in changing the lithospheric rheology and localising strain (e.g. Gernigon et al. this volume; Geoffroy 1998; Geoffroy et al. 2007; Gac and Geoffroy 2009; Ebinger et al. 2013). The nature of the crust can be ambiguous in highly thinned areas of "transitional" crust that appears to

show neither classic oceanic or continental crustal properties. Finally, terms such as 'continental suture' are difficult to define and can have complex, three-dimensional geometries and do not represent a simple lineament. Such a suture zone could reactivate, not where it appears at the surface, but where it is weakest at depth.
The imperfect fit of the Wilson Cycle concept to observations (e.g. Krabbendam 2001; Buiter and Torsvik 2014; Dalziel and Dewey 2018) shows that the process of opening an ocean is more complex than a simple 2D-unzipping of continental sutures.

3 What is structural inheritance?

Continents contain broad zones of active deformation that extend deep into their interiors (Gordon 1998; Nielsen et al. 2007, 2014; Şengör et al. 2018). Such non-rigid behaviour departs significantly from the original paradigm of rigid plate tectonics. It results from the presence, preservation and repeated deformation of crustal and mantle-lithospheric mechanical weaknesses (Thatcher 1995; Holdsworth et al. 2001). The buoyancy of continental crust means that it, and its underlying lithospheric mantle, are not subducted in the same way as oceanic crust. As a result, zones of pre-existing weakness are preserved in the continental lithosphere and can be rejuvenated many times during successive phases of deformation over geologic time (Sutton and Watson 1986). *Structural inheritance* is a property of the continental lithosphere that guides deformation along pre-existing rheological heterogeneities at all scales. When this occurs under a given stress regime, the resulting process is known as (structural) *rejuvenation.*

Rejuvenation (Figure 5a) includes (i) *reactivation*, the repeated focussing of deformation along discrete pre-existing structures, e.g., faults, shear zones or lithological contacts and (ii) *reworking*, the repeated focussing of metamorphism, ductile deformation, recrystallisation, metasomatism and magmatism into the same lithospheric volume. Reactivation is primarily controlled by the compositional and mechanical properties of pre-existing structures, whilst reworking is primarily influenced by the thermal history of the lithosphere (Holdsworth et al. 2001).

At shallow depths, brittle fracturing or frictional sliding occurs, with slip facilitated by low-friction minerals such as talc, serpentinite and smectite (e.g., Escartín et al. 2003; Moore et al. 2004; Schroeder and John 2004). The transition between brittle and ductile deformation in crystalline rocks is dependent on temperature, composition and strain-rate and typically occurs at crustal depths of 10-15 km (Figure 5b; Sibson 1977; Gueydan et al. 2014). Movement along deformation zones is characterised by diffusion-accommodated viscous creep in phyllosilicate-rich rocks in this depth range. In the viscous regime, deformation is typically plastic and distributed over broader, more diffuse zones (Holdsworth 2004; Jefferies et al. 2006; Imber et al. 2008) (Figure 5b), but strain localisation here is still widespread at different scales (Braun et al. 1999; Precigout et al. 2007).

It is important to emphasise that, although reactivation controlled by structural inheritance is widely recognised along the NE Atlantic margin, this process should not always be assumed to be the primary control on lithosphere-scale rifting. A coincidence in rift-related structural trends with those of older basement structures may be a good indicator for reactivation, but is not in itself actual proof (see discussions in Holdsworth et al. 1997; Roberts and Holdsworth 1999), especially when structures are mapped at depth in the offshore. The most conclusive test for inheritance in offshore rift systems is the recognition of reactivation in correlative onshore regions (e.g. Wilson et al. 2006; Peace et al. 2018b).

4 Structural inheritance and rejuvenation at different scales

The lithospheric and basin evolution of the CNAR was likely governed by a complex combination of rejuvenation of different inherited structures and fabrics with different scales and orientations, alongside other processes such as magmatism. Lithosphere-scale rejuvenation includes almost every conceivable process that affects lithospheric rheology, locally or as a whole. These include changes in crustal and lithospheric thickness, thermal state and composition, sedimentary basin processes (faulting, sedimentation) and the mechanical heterogeneities of metamorphic and intrusive fabrics (Dunbar and Sawyer 1989a; Krabbendam and Barr 2000; Nagel and Buck 2004; Yamasaki and Gernigon 2009; Tommasi et al. 2009; Huismans and Beaumont 2011; Brune et al. 2014; Manatschal et al. 2015; Tommasi and Vauchez 2015; Petersen and Schiffer 2016; Duretz et al. 2016).

4.1 Bulk lithosphere structure, composition and thermal history

Post-Archaean orogenic processes generally led to lithospheric volumes that are weaker and warmer compared to stable cratonic lithosphere (Cloetingh et al. 1995; Krabbendam and Barr 2000; Rey et al. 2001; Corti et al. 2007). This may not always be the case, as, Krabbendam (2001) hypothesise that orogens with low heat flow (and "cold" crustal geotherms) have strong lithosphere, impeding reactivation. Nevertheless, the alignment of new structures with old weaknesses is persuasive, and has historically led many authors to postulate that reactivation is a major factor in breakup (e.g. Dunbar and Sawyer 1989a).

Numerical modelling suggests that discrete pre-existing lithospheric heterogeneities localise strain and control rift distribution (Dunbar and Sawyer 1989b) and asymmetric conjugate margin geometries (Yamasaki and Gernigon 2009; Petersen and Schiffer 2016; Beniest et al. 2018). Therefore, rifts generally localise at the boundaries of lithospheric blocks of varying rheology (Pascal and Cloetingh 2002; Beniest et al. 2018). The relative strength between crust and mantle lithosphere is strongly influenced by crustal thickness and this also governs depth-dependent extension and thinning (Huismans and Beaumont 2011; Petersen and Schiffer 2016) (Figure 6). Thickened, warm and weak crust can undergo delocalised thinning, whilst the mantle lithosphere is more abruptly thinned (Buck 1991; Huismans and Beaumont 2011). Preferential thinning of mantle lithosphere leads to decompression melting of the asthenosphere which can occur while the crust remains intact (Petersen and Schiffer 2016) (Figure 6). Increasing obliquity to the extension direction and curvature of the zone of thickened crust produce more asymmetric and segmented rift zones (Van Wijk 2005; Corti et al. 2007). In contrast, a thinned crust with a shallow Moho prior to extension and/or longer periods of thermal relaxation (>30-50 Ma) can produce a cold and strong lithosphere, impeding rift localisation (Harry and Bowling 1999; van Wijk and Cloetingh 2002; Guan et al., 2019). If the mantle is weaker than the crust, it flows laterally whilst the crust is locally thinned, forming narrow necking zones and impeding pre-breakup melt generation (Petersen and Schiffer 2016) (Figure 6).

The lithosphere beneath stagnated rifts may cool and harden, leading to rift jumps away from the stronger lithosphere of the old rift, producing asymmetric continental margins (van Wijk and Cloetingh 2002; Naliboff and Buiter 2015). Such a process has been proposed to explain the formation of the volcanic margins in the NE Atlantic off-axis from previously thinned crust and failed rifts hosting Palaeozoic and Jurassic sedimentary basins (Gernigon et al. this volume; Guan et al. 2019). The zone of rheological contrast of such cooled/re-equilibrated rift zones and associated sedimentary infill may be reactivated during later episodes of extension, or may partition deformation (Odinsen et al. 2000; Frederiksen et al. 2001; Brune et al. 2017).

Armitage et al. (2010) demonstrated that thinned lithosphere from prior rift phases can enhance melt productivity in a subsequent rift phase.

Lithospheric delamination has been suggested to have a major impact on rift evolution and magmatism (Bird 1979; Kay and Kay 1993; Meissner and Mooney 1998; Elkins-Tanton 2005; Meier et al. 2016; Petersen et al. 2018). Şengör et al. (2018) suggested that rejuvenation of pre-existing structures may be linked to removal of the lithospheric mantle, which would weaken the entire remaining lithospheric column. Subsequently, extensive magmatism would inhibit thermal re-equilibration of the lithosphere and allow rejuvenation to continue for a long time. Liu et al. (2018) and Wang et al. (2018) propose models where a "Mid-Lithospheric Discontinuity" or the lower crust can act as a sub-horizontal weakness zone along which the lithosphere may delaminate.

4.2 Discrete lithospheric structures

Discrete structures include regional-scale features such as sutures, shear zones, igneous bodies and other large features found at depth within the lithosphere.

Pre-existing rheological heterogeneities such as suture, fault and shear zones, possibly incorporating preserved eclogite and hydrated peridotite within the continental lithosphere may influence rifting, location of breakup and margin architecture (e.g. Petersen and Schiffer 2016).

Pre-existing mafic or ultramafic magmatic rocks are known to increase crustal strength and viscosity (Burov 2011). For example, the development of continent-dipping bounding faults of SDRs at magma-rich passive margins requires increased lower crustal viscosities (Geoffroy et al. 2015). Syn-rift magmatic systems such as crustal intrusions (Ebinger and Casey 2001; Keir et al. 2006), crustal magma chambers (Geoffroy 1998; Doubre and Geoffroy 2003) or instabilities at the lithospheric thermal boundary layer (Geoffroy et al., 2007; Gac and Geoffroy, 2009) weaken the lithosphere and can accommodate and localise deformation (Buck and Karner 2004) at different lithospheric levels during the rifting process.

In the North Atlantic rifting and breakup-related magmatism was typically focussed in igneous centres. Some of those igneous centres are located along pre-existing inherited fault and shear zones (e.g., the Great Glen Fault) (Bott and Tuson 1973; Geoffroy et al. 2007; Gueydan et al. 2014). The spacing, location, size and magmatic budget of these igneous centres are governed by complex interactions between pre-existing discrete structures, pre-existing lithospheric thickness variations and mantle composition, as well as the timing and degree of melting (Gernigon et al. this volume; Gouiza and Paton 2019).

High-velocity lower crustal bodies (HVLCBs) are observed along most continental margins of the CNAR (Mjelde et al. 2008; Lundin and Doré 2011; Funck et al. 2016a) (Figure 7). Identifying the origin of HVLCBs is essential to understand extension and magmatism in rifts and passive margins. Many are associated with magmatic underplating or intrusions added to the lower continental crust during extension (Olafsson et al. 1992; Eldholm and Grue 1994; Ren et al. 1998; Mjelde et al. 2007b; White et al. 2008; Thybo and Artemieva 2013; Wrona et al. 2019). However, it is unclear to what extent such features are emplaced during rifting related to breakup. Some HVLCBs in the CNAR have been interpreted as metasomatised, metamorphosed or intruded mafic rocks in the uppermost mantle originating from Caledonian or older subduction and collision zones (Abramovitz and Thybo 2000; Christiansson et al. 2000; Gernigon et al. 2004, 2006; Ebbing et al. 2006; Wangen et al. 2011; Fichler et al. 2011; Mjelde et al. 2013; Nirrengarten et al. 2014; Schiffer et al. 2015a, 2016; Abdelmalak et al. 2017; Slagstad et al. 2018) (Figure 7). If the HVLCBs are deformed, pre-existing structures, they will likely have influenced and localised the rifting before breakup-related magmatism (Gernigon et al. 2004; Petersen and Schiffer 2016).

Salazar-Mora et al. (2018) showed that during rifting of an orogenic belt, initial reactivation usually occurs along pre-existing lithospheric-scale suture zones, whilst the amount of previous contraction governs the width of the reactivated crustal segment and its offset from the suture. Thus, pre-existing contractional shear zones are reactivated first and new shear zones form later. Intrusions in the upper crust may weaken the surrounding rock and control breakup localisation (Geoffroy et al. 1998). Increasing obliquity of crustal weak zones encourages increasingly diffuse rift zones, delaying lithospheric breakup (Brune et al. 2014). Heron et al. (2016; 2018) showed that reactivation of long-lasting intraplate "mantle scars" may lead to substantial intraplate deformation. Like many of the above studies, they also emphasised that mantle heterogeneities are usually favourably reactivated in comparison to crustal structures. This is because these are the load-bearing layers of the lithosphere (e.g. Holdsworth et al., 2001).

4.3 Pervasive lithospheric fabric

Small-scale compositional and rheological variations form fabrics in the crust and mantle and localise strain, forming complex patterns of crustal-scale, anastomosing shear bands, lithospheric boudinage structures, crustal rafts or continental ribbons in continental margins (Lister et al. 1986; Clerc et al. 2015; Jammes and Lavier 2016). Similarly, extension of a chemically heterogeneous, finely layered lithosphere leads to boudinage/necking of relatively strong layers causing intense structural softening as weaker layers become mechanically interconnected (Duretz et al. 2016).

Rifting and continental breakup may exploit anisotropies formed during previous phases of deformation in the lithospheric mantle (Vauchez and Nicolas 1991; Tommasi and Vauchez 2001; Misra 2016). Seismic anisotropy of the lithosphere may reflect mechanical anisotropy and is often, but not always, parallel to mountain/deformation belts (e.g. Vauchez et al. 1997; Tommasi and Vauchez 2001; Huang et al. 2006; Barruol et al. 2011). With some exceptions, the general trend of the fast direction of shear wave splitting along the North Atlantic margins is aligned with that of Caledonian-Variscan structures and deformation (e.g. Helffrich 1995; Barruol et al. 1997; Kreemer 2009; Wüstefeld et al. 2009; Darbyshire et al. 2015; Wang and Becker 2019).

Regional seismic tomography shows that present-day mantle anisotropy is generally aligned with late-Caledonian shear zones in the British Isles, the North Sea and southern Norway (GGF, WBF, HBF, MTFC, HFZ, Fig. 2,4), but oblique to those farther north (north of the Jan Mayen microplate complex) (Zhu and Tromp 2013). While breakup in the southern NE Atlantic followed the general Caledonian orogenic trends, breakup in the northern NE Atlantic (NE Greenland-NW Norway) followed an oblique, more easterly trend relative to the main Caledonian axis (defined as the central/median line between the orogenic fronts). This trend follows the late orogenic sinistral shear fabric of the NE Atlantic (Soper et al. 1992; Dewey and Strachan 2003), a fabric that was likely also reactivated during Late Caledonian extension (Figure 4) (Andersen et al. 1991; Dewey et al. 1993; Fossen 2010). This suggests that the mantle fabric and line of breakup in the north are to some extent related, while crustal fabric and breakup can be oblique.

A critical observation is that most of the rift systems that predated breakup (e.g. SW Barents Sea Basins, Danmarkshavn Basin, the Lofoten, Vøring, Møre basins, Faroe-Shetland basins, Hatton and Rockall basins (Tsikalas et al. 2012; Gaina et al. 2017; Stoker et al. 2017) largely followed the major orogenic NE-SW crustal trends (Figure 8). In Section 7.1 we propose that pre-breakup continental rift systems inherited the shallower, crustal fabric mainly, whilst the later breakup dominantly exploited the oblique, deeper, pervasive, mantle fabrics, controlled by a major change in stress field.

4.4 Crustal-basin scale concepts

At crustal-basin scale, deformation typically localises along weak zones such as pre-existing faults or shear zones. The size, geometry and interconnectivity of the discrete structures control the amount and magnitude of reactivation (Holdsworth 2004). Basin-scale structures exert a range of influences over later tectonic events, including strain localisation to control rift and fault nucleation, and also partitioning strain to potentially segment and block the propagation of rift related structures. But whether, and how, rifting is influenced depends on the type and geometry of the pre-existing structure and its relation to the imposed stress field. The orientation of the extensional stress field controls which older crustal structures reactivate at a given time, resulting eventually in co-linear/sub-parallel alignments between basins and older orogenic structural trends (Shannon 1991; Bartholomew et al. 1993; Doré et al. 1997; Roberts et al. 1999).

Pre-existing faults undergo varying degrees and styles of reactivation during later rift events (Bell et al. 2014; Whipp et al. 2014; Henstra et al. 2015; Deng et al. 2017b). The presence and reactivation of pre-existing basement structures, such as pervasive fabrics or discrete structures, can produce fault and rift geometries that depart from idealised geometries for orthogonal rift systems (Morley et al. 2004; Paton and Underhill 2004; Whipp et al. 2014). These effects may manifest as fault patterns oriented oblique to the regional stress field, and may also display complex internal transfer and linkage patterns (Morley et al. 2004; Bird et al. 2015; Bladon et al. 2015; Mortimer et al. 2016). In some instances, pre-existing structures may transfer strain across a rift from margin to axis as extension progresses (Morley et al. 2004; Bladon et al. 2015; Mortimer et al. 2016). Pre-existing structures can also act as stress guides that locally rotate the maximum horizontal stress in the overlying basin, controlling the trends of newly forming structures (Morley 2010; Whipp et al. 2014; Duffy et al. 2015; Reeve et al. 2015; Phillips et al. 2016). Oblique extension or transtension in the presence of pre-existing weak zones commonly leads to partitioning displacement into strike-slip and dip-slip fault components (De Paola et al. 2006; Wilson et al. 2006; Philippon et al. 2015; Kristensen et al. 2018).

Steeper dipping structures are preferentially reactivated under extensional stress compared to shallowly dipping structures (Bird et al. 2015; Phillips et al. 2016; Fazlikhani et al. 2017). Daly et al. (1989) show that gently dipping shear zones may be reactivated in a dip-slip manner whereas steeply dipping structures tend to display strike-slip reactivation. Structures at high angles to the regional stress direction (typically > 45°) are typically not reactivated (Henstra et al. 2015; Deng et al. 2017b; Henstra et al. 2017; Deng et al. 2018) and may be cross-cut by later faults (Duffy et al. 2015; Henstra et al. 2015; Phillips et al. 2016; Fazlikhani et al. 2017). Alternatively, they may inhibit fault propagation and segment rift basins (Doré et al. 1997; Fossen et al. 2014; Nixon et al. 2014).

During multiple phases of extension, pre-existing fault networks influence the development of later faults. Faults that reactivate pre-existing structures often quickly attain the length of the reactivated structure before undergoing displacement-dominated growth (Walsh et al. 2002; Whipp et al. 2014; Childs et al. 2017). The influence of pre-existing faults may be complicated by healing during burial following earlier rift phases, the combination of pre-existing fault orientations and the applied stress orientation, as well as lithospheric properties (Cowie et al. 2005; Baudon and Cartwright 2008; Henza et al. 2011; Bell et al. 2014; Whipp et al. 2014; Henstra et al. 2015, 2017; Claringbould et al. 2017). Complex fault geometries from multi-phase rifts have been documented in natural examples (Nixon et al. 2014; Duffy et al. 2015; Reeve et al. 2015; Rotevatn et al. 2018) and simulated in analogue models (Keep and McClay 1997; Corti et al. 2007; Henza et al. 2010, 2011; Henstra et al. 2015; Duffy et al. 2017) although in some instances complex, non-colinear fault networks may also arise in single-phase rifts due to a 3D stress field (Healy et al. 2015; Collanega et al. 2017; Gernigon et al. 2018).

The emplacement of igneous rocks may be controlled by pre-existing structures at the crustal (Peace et al. 2017) and intrusion scale (Peace et al. 2018c). Igneous complexes may subsequently favour the nucleation and formation of new shear zones (Neves et al. 1996) and can lead to spatial variations in deformation patterns within rift systems and basins (Woodcock and Underhill 1987; Buck 2006; Dineva et al. 2007; Magee et al. 2014, 2017; Phillips et al. 2017). Steeply dipping intrusions, such as dyke systems, may promote strain localisation in a similar way to basement faults and fabrics during rifting, introduce anisotropy and controlling the geometry and evolution of faults (Buck 2006; Ruch et al. 2016; Phillips et al. 2017). In contrast, sub-horizontal intrusions such as sills and laccoliths may produce more distributed strain patterns caused by uplift and outer arc extension in forced folds at sub-basin scales (Wilson et al. 2016; Magee et al. 2017).

Compression of previously formed rift basins typically leads to basin inversion (Stephenson et al. this volume; Buchanan and Buchanan 1995; Lowell 1995). During basin inversion, the geometry of the extensional faults, which may themselves be influenced by basement fabric, affects the style of inversion produced when reactivated under oblique convergence (Withjack et al. 2010; Kley 2018). On the NE Atlantic margins, the widespread Cenozoic inversion structures (Stephenson et al. this volume; Johnson et al. 2005; Doré et al. 2008; Pascal and Cloetingh 2009) also seem to track underlying extensional basin and lithospheric structure (Nielsen et al. 2014). This, in turn, was probably inherited indirectly from basement fabric (Kimbell et al. 2017; Reilly et al. 2017).

5 Pre-rift structural framework of the Circum-North Atlantic region

The main accretionary events predating CNAR breakup were the mid-Neoproterozoic Sveconorwegian-Grenvillian (Bingen et al. 2008b; Roberts and Slagstad 2015), the Neoproterozoic Timanian (Roberts and Siedlecka 2002; Gee and Pease 2004) and the Phanerozoic Caledonian (Roberts 2003; Gee et al. 2008) and Variscan orogenies (Matte 2001; Winchester et al. 2002; Franke 2006). These orogenies were in essence the expressions of two Wilson cycles: (i) the assembly and dispersal of the supercontinent Rodina in the Neoproterozoic leading to the formation of the Iapetus Ocean, followed by (ii) the renewed assembly and dispersal of Pangaea in the Phanerozoic and formation of the North Atlantic in the Cenozoic (Stampfli et al. 2013).

5.1 Archaean-Proterozoic cratons

The North American and East European cratons, the respective cores of the palaeocontinents Laurentia and Baltica (prior to their Caledonian suturing to become Laurussia (Roberts et al. 1999; Ziegler 2012), were formed from the Archaean through to the Proterozoic and consist of terranes of different age separated by networks of mobile belts. In Baltica, the main tectonic episodes were the Archaean Karelian and Lapland-Kola events in northern Scandinavia, the Palaeoproterozoic Svecofennian orogeny in central Scandinavia, and formation of the Trans-

Scandinavian Igneous Belt (TIB) in the late Palaeoproterozoic from southern Sweden to NW Norway (Gorbatschev and Bogdanova 1993; Balling 2000). Similarly, Laurentia depicts differently aged cratonic terranes and mobile belts (St-Onge et al. 2009). In the CNAR, these include the North Atlantic Craton, the Rae Craton and the Superior Craton, conjoined by the Palaeoprtoerozoic Nagsuqqotidian, Makkovik-Ketilidian, Rinkian and other orogens (St-Onge et al. 2009) (Fig. 2, 3).

5.2 The Grenville-Sveconorwegian Orogeny

The Grenville-Sveconorwegian fold belt evolved during the assembly of Rodinia in the late Mesoproterozoic (Li et al. 2008). The Grenville Orogen in NE North America includes the collision between Laurentia and Amazonia (1.09-1.02 Ga), marked by high-grade metamorphism (Hynes and Rivers 2010; Rivers 2015). The basement of southern Scandinavia was assembled by several events prior to the Sveconorwegian orogeny: the Gothian (1.64-1.52 Ga), Telemarkian (1.52-1.48 Ga) and Hallandian events (1.47-1.42) (Bingen et al. 2008a). The actual Sveconorwegian Orogen is characterised by terrane accretion events between 1.14 and 0.97 Ga arising from collision between Baltica and other continental fragments, followed by orogenic collapse at 0.9 Ga (Bingen et al. 2008a). Although the Sveconorwegian orogeny was largely coeval, and likely also spatially related to the Grenville Orogen, the precise connection between these orogens is unclear, as well as the regional configuration, especially of Baltica at this time (Bingen et al. 2008b; Slagstad et al. 2013, 2019; Cawood and Pisarevsky 2017). Sveconorwegian-aged deformation is also reported in the Arctic but any relationship to the main fold-belt is unclear (Lorenz et al, 2012). The latest Neoproterozoic Valhalla Orogeny has been proposed as an accretionary orogen along Laurentia's free margin (East Greenland) (Cawood et al. 2010; Spencer and Kirkland 2016).

5.3 The Timanian Orogeny

The Timanian fold-and-thrust belt records ocean-continent collisions along the northern margin of Baltica and the accretion of island arc complexes, terranes and microcontinents at ~0.62-0.55 Ga stretching from the Scandinavian Arctic to the Arctic Urals (Roberts and Siedlecka 2002; Gee and Pease 2004; Gee et al. 2006, 2008). The Trollfjorden-Komagelva Fault Zone is a major Timanian structure extending from the Urals across the Timan Range to northernmost Norway, where it was later reworked by the Caledonides (Gernigon and Brönner 2012; Gernigon et al. 2014, 2018; Klitzke et al. 2019). Neoproterozoic Timanian basement terranes, metasediments and volcanic sequences were drilled in the Pechora Basin (Roberts and Siedlecka 2002; Dovzhikova et al. 2004). The Timanian suture may be deeply buried in the central Barents Sea (Gernigon et al. 2018) possibly associated with high velocity-high density lower crustal rocks (Shulgin et al. 2018). Basement structures in the eastern and central Barents Sea show a persistent NW-SE oriented Timanian fabric throughout the region (Gee et al. 2006, 2008; Pease 2011; Klitzke et al. 2019). The Torellian orogeny on Svalbard may be a Timanian equivalent or prolongation (Majka et al. 2008).

5.4 The Caledonian Orogeny

Prior to opening of the North Atlantic Ocean, Europe, North America and Greenland comprised part of the most recent continental amalgamation, Laurasia, the northern constituent of Pangaea (reconstructions in Figure 3,4) (Cocks and Torsvik 2006, 2011; Lawver et al. 2011; Stampfli et al. 2013). As part of Laurasia, Laurussia was formed by closure of the Iapetus Ocean and Tornquist seaway, and collision of three palaeocontinents – Laurentia, Baltica and Avalonia – as well as smaller terranes, culminating in the Scandian phase of the Caledonian orogeny at 425-400 Ma (Soper and Woodcock 1990; Pharaoh 1999; McKerrow et al. 2000; Roberts 2003; Gee et al. 2008; Leslie et al. 2008). This was preceded by phases of arc-accretion in Norwegian, British and North American Caledonides in the late Cambrian-early Ordovician, i.e. the

Finnmarkian/Jämtlandian (Brueckner and Van Roermund 2007), Grampian (Dewey 2005) and Taconian stages (Karabinos et al. 1998), respectively. Laurasia's assembly was completed by accretion of the Siberian and Kazakhstan continental plates during the Uralian and Mongol-Okhotsk orogenies (Blakey 2008). Avalonia was the first of several large terranes released from Gondwana to dock against Baltica and Laurentia, opening the Rheic Ocean in the mid-Ordovician (Matte 2001). the docking of the peri-Gondwana terranes Armorica and Megumia to the south of Avalonia in the British Isles and Appalachians defines the Early Devonian Acadian stage after the complete closure of the Iapetus Ocean and docking of Avalonia (Murphy and Keppie 2005; Woodcock et al. 2007; Mendum 2012; Woodcock and Strachan 2012). In case of the Appalachians, other authors have proposed that the Acadian represents the actual docking of Avalonia during the closure of the Iapetus Ocean (Hatcher et al. 2010; Hibbard et al. 2010). In the Barents Sea the observed trends and seismic evidence suggest two branches of the Caledonian suture (Doré 1991; Gudlaugsson et al. 1998; Breivik et al. 2005; Gee et al. 2008; Gernigon et al. 2014), an NE-SW oriented branch and an N-S oriented branch parallel to the present-day western Barents margin towards Svalbard (Gudlaugsson et al. 1998; Breivik et al. 2002; Aarseth et al. 2017). Shortening in the Palaeozoic Ellesmerian fold belt in Svalbard, North Greenland and Arctic Canada was broadly contemporaneous with Caledonian deformation (Ziegler 1988; Gasser 2013; Gee 2015).

While the fundamental tectonic elements of the Caledonian orogeny are reasonably well understood, significant aspects of timing, deformation, polarity and number of subduction events are not resolved. Structural and tectonic relationships are complicated due to overlap and interaction of Caledonian structures with earlier structures (Roffeis and Corfu 2014). Ages of Caledonian metamorphism and intrusions in East Greenland and Scandinavia range from 500 to 360 Ma with early age populations (~500-422; Kalsbeek et al. 2008; Corfu et al. 2014), the main Scandian phase (~425 Ma; Dobrzhinetskaya et al. 1995; van Roermund and Drury 1998; Hacker et al. 2010) in Scandinavia and East Greenland, as well as young ages in NE Greenland (~360 Ma, (Gilotti et al. 2014), indicating complex and prolonged evolution (Gasser 2013; Corfu et al. 2014). These observations have led to departures from a simple model of only west-dipping Scandian subduction and collision. Other suggested models include additional early west-dipping (Brueckner and van Roermund 2004; Brueckner 2006) or east-dipping subduction events (Yoshinobu et al. 2002; Andréasson et al. 2003; Roberts 2003; Gee et al. 2008; Schiffer et al. 2014), possibly as a northward equivalent of the Grampian (Karabinos et al. 1998; van Staal et al. 2009) or Taconian phases (van Staal et al. 1998; Dewey 2005), and late intracratonic eastward underthrusting (Gilotti and McClelland 2011). Although the Caledonian orogeny between Greenland and Scandinavia can be approximated as a linear, "two-dimensional" orogen, complexities of Caledonian fabrics can be observed along the length of the orogen indicating a composite, non-orthogonal collision and subduction system (Fossen et al. 2008). The late Caledonian phases were dominated by the gravitational collapse of high, unstable topography, accompanied by lithospheric extension and possibly lithospheric delamination (Seranne 1992; Fossen et al. 2014; Gabrielsen et al. 2015) with major sinistral strike-slip along the Baltic and Laurentian margins (Harland 1969, 1971; Roberts 1983; Soper et al. 1992).

5.5 The Variscan Orogeny

Following consolidation of Laurasia in the Late Silurian-Early Devonian, the basement substructure of the southern CNAR was modified by the Variscan-Appalachian Orogeny, a major continent-continent collision to the south with Gondwana and peri-Gondwanan terranes and microcontinents (McKerrow et al. 2000; Franke 2006; Winchester et al. 2006; Kroner and Romer 2013). The episodic release of peri-Gondwana terranes was probably driven by back-arc spreading on the Gondwana margin (Stampfli and Borel 2002). These terranes successively docked against Laurasia to the north, each generating individual compressional pulses. While the orogenic evolution of the Appalachians is relatively well-defined (Hatcher et al. 2010), the situation is more complicated in the European Variscides, due to a more complex subduction history (Matte 2001). The Variscan Orogeny ended with collision between Gondwana and Laurasia in Late Carboniferous-Permian time (McKerrow et al. 2000; Matte 2001), forming a major fold belt, running E-W through southern Europe and NE-SW between eastern North America and NW Africa.

This collision involved dextral transpression and likely resulted in major orogen-parallel transform faults in the Appalachians (Hatcher 2002). Similarly, the Variscides of the Iberian Peninsula were bounded by the NW-trending Coimbra-Cordoba and Ossa-Morena shear zones in the south, likely connected to the North Iberia Fault. The Coimbra-Cordoba shear zone experienced at least 72 km of sinistral motion (Burg et al. 1981). A transform system may have continued from the North Iberia Fault along the proto-Flemish Cap and Goban Spur margins, through the proto-Labrador Sea, connecting with the Hudson Strait-Foxe Channel fault system (Lundin and Doré 2018).

6 Structural segmentation and inheritance in the CNAR

The CNAR margins are segmented in terms of crustal thickness, width, basin thickness, magmatism and presence of HVLCBs (Skogseid et al. 2000; Lundin and Doré 2011; Peron-Pinvidic et al. 2013; Funck et al. 2016b; Ady and Whittaker 2018; Lundin et al. 2018). A number of failed rift systems are present, including the North Sea Central and Viking grabens, the conjoined Møre and Vøring basins, the parallel Rockall and Hatton basins, plus the Porcupine, Orphan, Danmarkshaven and Bjørnøya basins (Ziegler 1992; Péron-Pinvidic and Manatschal 2010; Lundin and Doré 2011; Gernigon et al. 2014). This rift network on the continental shelves may be linked to pre-existing lithospheric-scale structures, lineaments and terranes (Doré et al. 1997; Chenin et al. 2015; Gaina et al. 2017; Schiffer et al. 2018).

A key observation in the northern NE Atlantic is that in some areas, the late Caledonian shear zones, primarily recognised in the onshore, are largely parallel to the breakup trend, while in other areas there is a distinct obliquity of the breakup axis with earlier rift basins that follow Caledonian trends (Fig. 8). This obliquity likely relates to interaction between different sets and depths of pre-existing structures and varying extensional stress-fields.

The relationship between the Caledonian Orogen and the CNAR can be described in the context of five primary segments (Fig. 1):

(1) In the northern section between East Greenland and Norway, many of the late Palaeozoic-Early Cretaceous rift basins (Møre, Vøring, Lofoten-Vesterålen, Danmarkshavn basins, and possibly the Thetis Basin) follow the mapped NE-SW Caledonian trends, while latest Cretaceous- earliest Cenozoic rifting and breakup is clockwise oblique by ~20-30º (Figures 4,8). North-south Mid-Late Jurassic faulting, as expressed in the Halten Terrace, appears to be strongly discordant to this pattern, although a link to duplex systems between Caledonian shears was suggested by Doré et al., (1997).

(2) In the southern NE Atlantic, between SE Greenland and the British Isles, the Cretaceous-Jurassic Rockall and Hatton basins and the axis of breakup all follow the general Caledonian trend. However, breakup produced highly asymmetric margins lying 500 km or more west of the Caledonian front, and cutting through cratonic lithosphere (Figures 3, 4, 8).

(3) The Greenland-Iceland-Faroe Ridge (GIFR) separates segments 1 and 2. Formation of this ridge is discussed in detail by Foulger et al. (this volume). Relative movements between the Reykjanes Ridge to the south and the abandoned Aegir Ridge, and active Kolbeinsey Ridge to

the north must have been accommodated along the GIFR. Additionally, the GIFR formed where the North Atlantic rift crosscut the western Caledonian front and along the southern margin of the Rae Craton.

(4) The North Sea experienced rift phases in the Permian-Triassic, Late Jurassic and Early Cretaceous in the area of the Iapetus-Thor Suture triple junction and the Danish-German-Polish Caledonides. The physiography of the later rift is dominated by the northern Viking Graben, the western Moray Firth Graben and the Central Graben (Færseth et al. 1995). As indicated in point (1), the dominant N-S faulting of the northern North Sea, generally discordant to the Caledonian trends, is probably attributable to rift propagation from the SE (e.g. Figs. 4 and 8).

(5) The Labrador Sea and Baffin Bay are two ocean basins formed during the Palaeogene by a now-extinct spreading system, with the Davis Strait separating and accommodating relative motions between them. Rifting and continental breakup both cross-cut and ran sub-parallel to crustal terrane boundaries (Figs. 2,3).

The principal rift architectures and possible relations to inherited fabrics in these segments are as follows:

Segment 1 – Norway-Greenland margins

North of the GIFR, the conjugate margins of Norway and Greenland are asymmetric, with structural variations along strike, as a result of the oblique breakup axis (Fig. 8) (Gernigon et al. this volume). The narrow continental shelf in central East Greenland contrasts strongly with the wide conjugate Vøring margin. Further north, the NE Greenland shelf is much wider than its conjugate Lofoten margin.

The Mid-Norwegian margin is magma-rich with thick NE-SW trending sedimentary basins and highs. This NE-SW trend is attributed to Caledonian and Precambrian inheritance (Bergh et al. 2007; Maystrenko et al. 2017). The margin is divided into the Møre, Vøring and Lofoten-Vesterålen margins from south to north (Gernigon et al. this volume; Lundin and Doré 1997; Brekke 2000; Mosar 2003). This segmentation is related to margin-perpendicular transfer zones, the Jan Mayen Lineament/Corridor between the Møre and Vøring margins (Eldholm et al. 2002) and the Bivrost Lineament separating the Vøring from the Lofoten-Versterålen margin (Blystad 1995). These lineaments are expressed through offsets in basin axes, inferred by some authors to reflect Caledonian or Precambrian basement fabrics (Doré et al. 1997, 1999) or Late Jurassic–Early Cretaceous extensional structures (Eldholm et al. 2002).

As indicated earlier, Jurassic faulting, as expressed in the Halten Terrace (e.g. Blystad et al., 1995), is approximately N-S and strongly discordant to both the Caledonian trends, later (Early Cretaceous) basin formation and Cenozoic breakup (Fig. 8). The N-S faulting here and in the North Sea may represent an attempt by Tethys, the dominant oceanic domain at the time, to propagate through the North Sea into what is now the Norwegian Sea (compare Figs. 4 and 8).

In the Møre and Vøring segments, deep, inherited HVLCBs of Caledonian and/or Precambrian age controlled rifting (Gernigon et al. 2003, 2004; Abdelmalak et al. 2017; Maystrenko et al. 2017; Zastrozhnov et al. 2018). In the inner Møre margin, basin architecture follows the trends of late-Caledonian shear zones bordering the margin to the south, specifically the Møre-Trøndelag Fault Complex (Grunnaleite and Gabrielsen 1995; Hurich 1996; Jongepier et al. 1996; Doré et al. 1997; Nasuti et al. 2011; Theissen-Krah et al. 2017) suggesting a genetic connection. The dominant NE-SW-trending Caledonian thrust sheets in the Lofoten-Vesterålen margin interact with the Palaeoproterozoic NW-SE, margin-perpendicular Bothnian-Senja Fault Complex (Bergh et al. 2007), which appears to run into the younger offshore Senja

Fracture Zone, forming the Barents Sea's western transform boundary (Henkel 1991; Doré et al. 1997).

With the exception of the Wandel Sea Basin, the conjugate East Greenland margin is characterised by approximately N-S trending basins. Extensional faults appear to relate to Caledonian fabrics (Henriksen 2003) (Figure 9). Late Devonian-Early Carboniferous shear zones formed during Caledonian collapse (Surlyk 1990; Price et al. 1997; Parsons et al. 2017; Rotevatn et al. 2018), accompanied and possibly facilitated by major strike-slip deformation (Dewey and Strachan 2003), which may have reactivated older, pre-Caledonian shear zones related to the opening of the Iapetus Ocean (Soper and Higgins 1993). Faulting in the Triassic-Cretaceous East Greenland rift system was episodic, with multiple stages of reactivation culminating in the final separation of the JMMC (Surlyk 1990; Stemmerik et al. 1991; Hartz and Andresen 1995; Seidler et al. 2004; Parsons et al. 2017; Rotevatn et al. 2018). The East Greenland rift system is segmented by right-stepping NW-SE transfer zones (Fossen et al. 2017; Rotevatn et al. 2018). These offsets are thought to be related to reactivation of a NW-SE Proterozoic fabric (Andresen et al. 1998; White and Hodges 2002; Guarnieri 2015; Rotevatn et al. 2018).

The JMMC is located between the central East Greenland margin and the Møre margin (Gaina et al. 2009; Gernigon et al. 2015; Blischke et al. 2017, 2019; Polteau et al. 2018; Schiffer et al. 2018). The nature and formation of the JMMC remains enigmatic but its location and geographic relation to the GIFR and known Caledonian structures suggests inheritance control (Gernigon et al. this volume; Schiffer et al. 2015b, 2018) (see section 7.3). A lower crustal-upper mantle fabric, exemplified by a proposed N-S Caledonian (or pre-Caledonian) fossil suture zone (Schiffer et al. 2015b; Petersen and Schiffer 2016) may also have influenced rifting. However, direct geological or geophysical evidence for reactivation of older structures remains sparse.

The Wandel Sea Basin formed by transtension or extension during the mid-Cretaceous, and was modified by Palaeocene-Eocene N-S compression (Svennevig et al. 2016), synchronous with formation of the West Spitsbergen fold-and-thrust belt. Local structural trends (~NW-SE) closely mimic the conjugate Bothnia-Senja Fault Complex and Senja Fracture Zone. The Wandel Sea Basin is thought to have experienced multiple phases of reactivation of earlier rift structures (Guarnieri 2015).

Segment 2 – SE Greenland-Rockall-Hatton margins

The NE Atlantic south of the GIFR broke up parallel to Caledonian trends and structures, but ~500 km west of the Caledonian front through the Laurentian basement of the Rockall-Hatton margin. The margins in this segment are highly asymmetric. The Hatton margin comprises thinner and narrower SDRs and HVLCBs compared to SE Greenland (Planke and Alvestad 1999; Hopper et al. 2003). The SE Greenland continental shelf is straight and narrow, whilst the Rockall-Hatton margin shelf is extremely wide and formed during Jurassic-Cretaceous lithospheric thinning (Stoker et al. 2017).

The Rockall-Hatton margin contains two large failed rift basins (the highly extended, deep Rockall Basin and the less extended, shallower Hatton Basin) bounded by major marginal highs (Hatton High, Rockall Bank) (Morewood et al. 2005). The highs are underlain by crustal blocks up to 30 km thick (Funck et al. 2016a). The Rockall Basin has crustal thicknesses of <10 km beneath up to 5 km of sediments (Funck et al. 2016a) and is underlain by HVLC or hydrated mantle peridotite (Roberts 1975; Roberts et al. 1988, 2018; Makris et al. 1991; Shannon et al. 1999; Klingelhöfer et al. 2005; Morewood et al. 2005; Funck et al. 2016a). Reactivation of NNE–SSW to NE–SW, margin-parallel Caledonian and pre-Caledonian basement lineaments seems to have led to the initial localisation and segmentation of the Rockall-Hatton shelf.

Furthermore, the shelf is transected by NW-trending continental lineaments/transfer zones (Figure 9) (Rumph et al. 1993; Kimbell et al. 2005a; Stoker et al. 2017). Some of these lineaments link to faults onshore Ireland (e.g., SHL), others are associated with COB offsets of the Hatton-Rockall shelf (e.g., SHL, ADL) or are correlated with sedimentary basins (e.g., ADL, WTL, JF), and some may have guided magmatic intrusions or be related to oceanic fractures or accommodation zones in the Iceland Basin (e.g., CL) (Kimbell et al. 2005a; Naylor and Shannon 2005; Štolfová and Shannon 2009).

During Cenozoic compression/transpression, some transfer zones became the loci for inversion (Doré and Lundin 1996; Doré et al. 1999; Eldholm et al. 2002; Kimbell et al. 2005a; Tuitt et al. 2010). These are interpreted as rejuvenated Precambrian terrane boundaries or shear zones that had previously impeded rift propagation (Shannon et al. 1995, 1999; Kimbell et al. 2005a; Ritchie et al. 2008; Štolfová and Shannon 2009), thereby compartmentalising rift evolution in the Rockall Basin (Rumph et al. 1993; Kimbell et al. 2005a; Stoker et al. 2017). Such pre-existing, margin-perpendicular terrane boundaries between different blocks may also explain why the lithosphere beneath Rockall did not break, while rifting was transferred outboard to a weaker section (Johnson et al. 2005; Elliott and Parson 2008). At the southern margin of the Rockall Basin, a probable connection between the Charlie Gibbs Fracture Zone and the Iapetus Suture, suggests a further reactivation of a pre-existing Caledonian lithospheric feature (Shannon et al. 1994; Buiter and Torsvik 2014; Ady and Whittaker 2018) (Figure 2).

The poorly known, narrow margin in SE Greenland comprises SDRs, HVLC bodies, and igneous centres and intrusions (Dahl-Jensen et al. 1998; Korenaga et al. 2000; Callot et al. 2001; Klausen and Larsen 2002; Hopper et al. 2003). Palaeoproterozoic discontinuities in SE Greenland were reactivated as left-lateral shear zones prior to breakup and the margin was inverted during the Eocene or later (Guarnieri 2015). The highly asymmetric line of Cenozoic breakup, outboard of the Hatton Basin and close to the SE Greenland coast (e.g. Figs. 4, 8 & 9), is a curious feature that appears to have formed without significant observable initial rifting. Because data is sparse, it is not possible to make any definite connection with older weaknesses. Speculatively, both the trend and straightness of this margin segment suggests a connection with the Late Caledonian shear fabric, as exemplified by faults such as the Møre-Trøndelag Fault Complex (e.g. Fig. 2). This line forms the shortest path from the Aegir Ridge to the Labrador Sea, and may have been created or exploited by dextral strike-slip associated with Labrador Sea opening (Lundin and Doré 2018) or high geopotential energy associated with th forming ridge triple junction located south of Greenland at that time (Kristoffersen and Talwani 1977; Roest and Srivastava 1989; Guan et al. 2019). An alternative hypothesis is that after Early Cretaceous rifting, the lithosphere in the Rockall-Hatton shelf re-equilibrated, cooled and strengthened (Guan et al. 2019), thereby leaving the SE Greenland shelf as the weakest pathway for breakup due to its thick, warm crust (45-55 km) and weak lithosphere.

Segment 3 – The Greenland-Iceland-Faroe Ridge and adjacent margins

The GIFR forms a WNW-ESE ridge spanning the NE Atlantic from central East Greenland to the Faroe-Shetland Basin (Foulger et al. 2019). The GIFR has anomalously high topography with typically 20-30 km thick crust (Foulger et al. 2003; Fedorova et al. 2005; Torsvik et al. 2015; Funck et al. 2016b; Haase et al. 2016), which thickens to 40 km beneath the central Iceland Plateau (Darbyshire et al. 2000; Du and Foulger 2001; Kaban et al. 2002; Gudmundsson 2003; Foulger et al. 2003; Fedorova et al. 2005). Thick basaltic lava flows cover the ridge (Horni et al. 2017; Hjartarson et al. 2017). The origin, structure and composition of the lithosphere beneath the GIFR remain poorly understood but there is significant evidence for a component of continental crust (Foulger 2006; Torsvik et al. 2015; Schiffer et al. 2018; Petersen et al. 2018; Foulger et al. 2019). The role of structural inheritance here is unknown, but the common location and orientation of the GIFR and the intersection of the North Atlantic rift axis with the Caledonian orogenic front suggest a link (Foulger and Anderson 2005; Schiffer et al. 2015b, 2018; Foulger et al. 2019). In addition, the recent recognition of Faroe-Shetland basement terrane immediately north of Scotland and the correlation of its southern boundary with that of the Rae Craton in Greenland (Holdsworth et al. 2019) mean that the southern margin of the GIFR follows this ancient terrane boundary.

The central East Greenland margin appears to be structurally and magmatically segmented by margin-perpendicular Precambrian structures that accommodated transform motion and localised intrusions (Karson and Brooks 1999). This segmentation was controlled by local magmatic centres, from which magma flow was guided and transfer zones defined (Callot et al. 2001; Klausen and Larsen 2002; Callot and Geoffroy 2004). Tegner et al. (2008) linked some of these tectonic lineaments to failed rifts, localised magmatism and breakup between central East Greenland and the JMMC.

The Faroe–Shetland margin consists of basins and highs, formed from the Late Palaeozoic to early Cenozoic plate breakup, followed by syn- to post-breakup magmatism, compressional tectonics and differential uplift and subsidence (Doré et al. 1999; Roberts et al. 1999; Johnson et al. 2005; Ritchie et al. 2008, 2011; Fletcher et al. 2013; Stoker 2016; Stoker et al. 2017, 2018). N–S to NE–SW and ESE–WSW to SE–NW structural trends follow regional fabrics observed in onshore basement rocks (Doré et al. 1997; Wilson et al. 2010). Many Devonian to Jurassic rifts exhibit Caledonian structural inheritance with a generally NNE trend such as the Outer Hebrides/Minch fault zones (Imber et al. 2001) (Fig. 9), faults within the West Orkney Basin (Bird et al. 2015) and the northeastern Faroe-Shetland Basin (Lamers and Carmichael 1999; Ritchie et al. 2011; Stoker et al. 2017) (Fig. 9).

In contrast, some lineaments in the Faroe Shetland Basin, including the southern boundary of the basin (Judd Fault), have a NW-SE orientation. Similar to the Rockall-Hatton margin, this structural trend is pre-Caledonian and may have created the transfer zones that compartmentalised the basin during the Mesozoic and early Palaeogene (Ritchie et al. 2011), although these features are not ubiquitous (Moy and Imber 2009). In the southern part of the basin, a W-to-NW trend prevails, including the Wyville-Thomson Lineament (Fig. 9), which reactivated during the Palaeocene (Kimbell et al. 2005a; Lundin and Doré 2005; Ziska and Varming 2008). Compressional structures formed in the Late Cretaceous (Booth et al. 1993; Grant et al. 1999; Stoker 2016) have been attributed to strike-slip tectonics linked to a shear margin (proto-plate boundary) separating Faroe–Shetland and SE Greenland, which reactivated old lineaments prior to breakup (Roberts et al. 1999; Guarnieri 2015; Stoker et al. 2018). Further compressional folding and differential uplift events occurred during the Eocene to early Neogene (Johnson et al. 2005; Stoker et al. 2005; Ritchie et al. 2008).

Segment 4 – North Sea & Tornquist Zone

The North Sea formed within basement that had been influenced by the Caledonian orogeny and Devonian orogenic collapse (Coward 1990; Andersen 1998; McKerrow et al. 2000; Fossen and Hurich 2005; Fossen 2010), with subsequent generally E-W extension beginning in the Permian-Triassic (Ziegler 1992; Fossen and Dunlap 1999; Frederiksen et al. 2001; Coward et al. 2003) and E-W to NW-SE extension in the latest Jurassic to Early Cretaceous (Brun and Tron 1993; Underhill and Partington 1993; Færseth 1996; Frederiksen et al. 2001; Coward et al. 2003; Arfai et al. 2014; Bell et al. 2014; Duffy et al. 2015; Deng et al. 2017a). Its development was also variably influenced by Permo-Carboniferous rifting and magmatism (Glennie et al. 2003; Heeremans and Faleide 2004; Neumann et al. 2004; Wilson et al. 2004), and far-field Alpine compression combined with ridge-push and gravitational forces from the high topography in Norway in the Late Cretaceous and Eocene (Biddle and Rudolph 1988; Cartwright 1989; Nielsen et al. 2005, 2007; Pascal and Cloetingh 2009; Jackson et al. 2013).

The North Sea region exhibits a range of upper mantle fabrics of Precambrian to Devonian age (Klemperer and Hurich 1990; Blundell et al. 1991; Abramovitz and Thybo 2000; Balling 2000; Fossen et al. 2014). Large-scale Moho and upper mantle shear zones originating from Devonian extension are imaged in the Norwegian North Sea (Fossen et al. 2014; Gabrielsen et al. 2015). HVLCBs in the southwest (Abramovitz and Thybo 2000) and NW (Christiansson et al. 2000) of the North Sea, and lower crustal fabrics (Klemperer et al. 1990) in the vicinity of the Iapetus suture offshore NW England are attributed to Caledonian collision and may have exerted structural control on the development of the rifts. Caledonian or Variscan upper mantle fabric along the eastern British coastline (Blundell et al. 1991) may also have influenced rifting. Pre-Caledonian dipping structures in the upper mantle were identified in the Skagerrak Sea between Norway and Denmark (Lie et al. 1990).

Structural inheritance within the North Sea seems to be related to N-S to NE-SW oriented Caledonian and Devonian lineaments (Bartholomew et al. 1993; Glennie 1998; Fossen 2010), along with the NW-SE trend of the Tornquist Zone in the south (Pegrum 1984; Bartholomew et al. 1993; Mogensen 1994). Caledonian nappes and late-Caledonian strike-slip shear zones in Norway (Andersen and Jamtveit 1990; Fossen 1992; Fossen and Dunlap 1998; Vetti and Fossen 2012; Fossen et al. 2017) and Scotland (Stewart et al. 1997, 1999) extend offshore beneath the North Sea rift (Bird et al. 2015; Reeve et al. 2015; Phillips et al. 2016; Fazlikhani et al. 2017). These were reactivated during later tectonic events (Phillips et al. 2016; Fazlikhani et al. 2017; Rotevatn et al. 2018) and exerted strong control on the Permo-Triassic structural development of the North Sea (Færseth et al. 1995; Færseth 1996; Lepercq and Gaulier 1996; Phillips et al. 2016; Fazlikhani et al. 2017).

The lithosphere-scale Tornquist Zone (TZ) spans Central Europe from SE to NW and extends across the Central North Sea (Figs. 2, 9), marking a major change in lithospheric and crustal thickness between Baltica to the NE and younger lithosphere to the SW (Berthelsen 1998; Pharaoh 1999; Cotte and Pedersen 2002; Babuška and Plomerová 2004; Janutyte et al. 2015; Mazur et al. 2015; Hejrani et al. 2015; Köhler et al. 2015). The TZ is associated with a series of NW-SE oriented crustal rift systems which have been periodically reactivated (Pegrum 1984; Berthelsen 1998; Mazur et al. 2015; Phillips et al. 2018). The TZ accommodated major late-Cretaceous compression associated with far-field stresses imposed by the Alpine orogeny (Berthelsen 1998; Nielsen et al. 2005, 2007; Jackson et al. 2013; Phillips et al. 2018).

At the rift scale, lithospheric thinning from Permian-Triassic, Carboniferous-Permian and Devonian extension, strongly influenced the Late Jurassic-Early Cretaceous rift in the North Sea (Walsh et al. 2002; Whipp et al. 2014; Duffy et al. 2015; Henstra et al. 2015; Reeve et al. 2015; Childs et al. 2017; Deng et al. 2017a). Thinning localised the thermal perturbation during later extension, resulting in a narrower and more localised rift focussed in the Viking Graben (Odinsen et al. 2000; Cowie et al. 2005) which, as indicated earlier, probably represented the main marine conduit between the Tethyan ocean and the proto-Norwegian Sea.

Segment 5 – Labrador Sea, Baffin Bay & Davis Strait

The Labrador Sea and Baffin Bay form an extinct early Cenozoic spreading system, with the Ungava Fault Zone running through the Davis Strait separating the two ocean basins (Figure 10). The basins formed by two-phase divergence between Greenland and North America (Chalmers and Pulvertaft 2001; Hosseinpour et al. 2013). A first phase of NE-SW extension started in the Early Cretaceous and culminated in Palaeocene continental breakup in the Labrador Sea (Srivastava and Keen 1995; Chalmers and Laursen 1995; Larsen et al. 2009; Abdelmalak et al. 2012, 2018; Pinet et al. 2013; Jones et al. 2017). Mesozoic-Early Cenozoic faulting was controlled by reactivation of pre-existing structures (Peace et al. 2018a, b). A second phase of NNE-SSW extension caused oblique spreading from the late Palaeocene (C25) to late Eocene (Roest and Srivastava 1989; Abdelmalak et al. 2012), which ceased at about 36 Ma (Roest and Srivastava 1989). The continental Davis Strait underwent sinistral transtension, but not breakup during the first stage (Wilson et al. 2006; Suckro et al. 2013; Peace et al. 2018b), followed by sinistral transpression during the second stage (Geoffroy et al. 2001; Suckro et al. 2013).

The Labrador Sea and Baffin Bay formed perpendicular to many lithospheric-scale Precambrian structures and fabrics, perhaps suggesting limited basement inheritance (Figure 2). However, purely based on similar trends of pre-existing structures with the Labrador Sea and Baffin Bay it is apparent that exceptions may exist. Direct evidence that any of these pre-existing structures did reactivate and guided the formation of the Labrador Sea and Baffin Bay is lacking however. For example, the Palaeoproterozoic Rinkian Orogen along the West Greenland margin of Baffin Bay (Grocott and McCaffrey 2017) and Precambrian normal faults (McWhae 1981) and strike-slip faults (van Gool et al. 2002; St-Onge et al. 2009) are sub-parallel to the Labrador Sea margin (Fig. 2). Early Cretaceous sinistral transform motion (Lundin and Doré 2018) and/or Jurassic mafic dyke swarms (Watt 1969; Larsen et al. 2009; Peace et al. 2016) could also have played a role in strain localisation in the Labrador Sea. Additionally, Neoproterozoic and Palaeoproterozoic dyke swarms in West Greenland and Baffin Island are parallel to sub-parallel to coastlines and continental margins of Baffin Bay and the Labrador Sea. In particular the Late Palaeoproterozoic-Early Mesoproterozoic Melville Bugt dyke swarm is strikingly parallel to the Baffin Bay continental margins (Buchan and Ernst 2006b). Klausen and Nilsson (2018) proposed a continuation of this dyke swarm through southern Greenland. Similarly, the Palaeoproterozoic BN-1 dyke swarm in SW Greenland is parallel to Labrador Sea breakup (Ernst and Buchan 2004). The Neoproterozoic Franklin-Thule dyke swarm is sub-parallel to the Baffin Bay continental margins on the Greenland side, but largely parallel to breakup on Baffin Island (Buchan and Ernst 2006a). Direct reactivation of these dykes or lithospheric rheological anisotropies reworked during dyke emplacement may have facilitated or guided rifting and breakup in Baffin Bay and Labrador Sea.

The Ungava Fault Zone in the Davis Strait is a major structural discontinuity (Geoffroy et al. 2001; Peace et al. 2017, 2018b; Abdelmalak et al. 2018) that may be related to Proterozoic basement structures and mantle scars (Geoffroy et al. 2001; Peace et al. 2018b; Heron et al. 2019). For example, the Palaeoproterozoic Torngat-Nagssugtoqidian orogenic belt (van Gool et al. 2002; Grocott and McCaffrey 2017) could have formed a rheological barrier, preserving thicker, continental-affinity crust and lithosphere in the Davis Strait (Heron et al. 2019). The HVLC underlying Davis Strait (Funck et al. 2007, 2012) could represent remnants of pre-existing metamorphosed or metasomatised crust or mantle (Petersen and Schiffer 2016; Peace et al. 2017).

The Labrador Sea and Baffin Bay margins are subdivided into magma-rich and magma-poor segments by major lithospheric structures, such as the Upernavik Escarpment in Baffin Bay (Chauvet et al., 2019) and the Grenville Front or the Ketillidian Mobile Belt in the Labrador Sea (Keen et al. 2018; Gouiza and Paton 2019). The poorly defined onshore continuation of the Upernavik Escarpment trends parallel to the Precambrian crustal fabric. A potential interplay between lithospheric inheritance and magmatism in the NW Atlantic has been proposed (Foley 1989; Larsen et al. 1992; Tappe et al. 2007; Peace et al. 2017). Excessive melting along the Davis Strait may also be related to older lithospheric structures (Larsen et al. 1992; Koopmann et al. 2014; Peace et al. 2017). Clarke and Beutel (this volume) link the Davis Strait Palaeogene picrites to sudden rupture of the thick Nagssugtoquidian lithosphere during breakup.

The conjugate margins of the Labrador Sea and Baffin Bay display significant structural and magmatic asymmetry (Chalmers and Pulvertaft 2001; Funck et al. 2012; Suckro et al. 2012; Welford and Hall 2013; Peace et al. 2016; Keen et al. 2017; Welford et al. 2018; Chauvet et al. 2019). This asymmetry could indicate that the Greenland lithosphere was weaker prior to rifting compared to the conjugate Labrador margin (Welford and Hall 2013). It could also be the consequence of strain migration associated with hyperextension (Brune et al. 2014) and, in the southern Baffin Bay, to the usual development of VPMs away from previous amagmatic rift systems due to strain hardening (Guan et al. 2019). Major shear zones, faults and basement structures onshore (Figure 10) may have controlled the fracture zones, structural divisions and basin architecture offshore (Welford and Hall 2013; Jauer et al. 2014; Peace et al. 2018b).

Moho topography seen in northern Baffin Bay may result from reactivation of large-scale pre-existing structures (Jackson and Reid 1994). Approximately N-S faults produced during Cretaceous rifting were reactivated during the Palaeogene deformation phase, when Greenland moved north relative to North America (Gregersen et al. 2016), causing the Eurekan Orogeny (Oakey and Chalmers 2012) during which Palaeozoic and Proterozoic structures were reactivated (Piepjohn et al. 2016; Schiffer and Stephenson 2017; Stephenson et al. 2017).

7 Discussion

7.1 Rifting, segmentation and breakup in the CNAR

Our review suggests that in the NE Atlantic *(segments 1-4)*, many of the late Palaeozoic to Cretaceous rift systems follow the trend of Caledonian structures, particularly the NE-SW-oriented sub-vertical, orogen-parallel sinistral strike-slip faults formed during the Silurian-Devonian (e.g. GGF-WBF, HBF, SUF, MTFC; Figures 3,9). There are, however, some exceptions such as the noted obliquity of the N-S Triassic-Jurassic rift trend expressed in (for example) the Viking Graben and Halten Terrace. This may have reactivated duplex structures formed during the late Caledonian (Doré et al., 1997) but could simply represent a newly created trend resulting from northwards Tethyan propagation (Figs. 4 & 8). In many cases NW-SE to WNW-ESE lineaments and transfer zones (Figure 9) further partitioned the structure and evolution of the NE Atlantic margins (Doré et al. 1997, 1999; Kimbell et al. 2005b). Some of these lineaments had pre-Caledonian history while others formed during the development of post-Caledonian basins. Of significance is the relation between continental transfer zones and oceanic fracture zones. In some cases, continental lineaments pass laterally into oceanic transfer faults (Figure 9). Many other margin-perpendicular lineaments are expressed by offsets in sedimentary basin architecture, but direct evidence of strike-slip motion is often lacking. A connection between continental transfer faults and oceanic fracture zones in the Vøring margin (Tsikalas et al. 2002; Mjelde et al. 2005) is questioned in more recent studies of modern magnetic data (Olesen et al. 2007).

A key observation in the northernmost NE Atlantic (essentially the Vøring margin) is the obliquity of its breakup axis with earlier rift basins and the Caledonian (surface) trend in the northern part (*segment 1*), compared to other areas where Caledonian structures appear to be more parallel to breakup. The late-Caledonian shear zones are strikingly parallel to the line of breakup (Figure 4 & 8) suggesting a causal link. As we explain below, a primary reason for this obliquity may lie in the existence of lithospheric layers in which differently oriented pre-existing fabrics rejuvenate at different times in response to changes in the regional stress field. This may be additionally linked to the magmatic development, as the extent of the Cenozoic pre- and syn-rift magmatism of the NAIP is also generally parallel to the final line of breakup, which may have been "perforated" by magmatic intrusions (Gernigon et al., this volume), and/or strike-slip deformation (Lundin & Doré, 2018).

Numerical modelling suggests that the dominating weaknesses may lie within the mantle lithosphere, rather than the crust (e.g. Heron et al. 2016) (see also section 2.2). We hypothesise that at the onset of rifting, the crustal and mantle lithospheric fabrics were oblique to one another. There existed an older, orogen-parallel, brittle crustal fabric and an oblique, younger, upper mantle shear fabric. Rifting in the North Atlantic experienced phases of varying stress orientations that rejuvenated either the less dominant, shallow crustal fabric or the dominant mantle fabric.

We propose a scenario for the NE Atlantic rifting and breakup as follows:

- The Caledonides formed as a notably linear orogen in the NE Atlantic resulting in a predominantly orogen-parallel fabric of crustal blocks, terranes, nappes and thrust faults (Figure 4, 8). At this time, lithospheric mantle fabric was parallel to the brittle crustal features. During late Caledonian sinistral transpression (Soper et al. 1992) the pre-existing discrete, brittle crustal fabric was reactivated as strike-slip faults, preserving their original orogen-parallel orientation. In contrast, the pervasive ductile lower crustal-upper mantle fabric was reworked and reoriented to ENE-WSW (rotated 20-30° clockwise about the orogenic axis).

- Devonian orogenic collapse was driven by body forces created by the high Caledonian topography mainly perpendicular to the axis of the mountain range (England and Houseman 1986; Molnar et al. 1993; Schiffer and Nielsen 2016), however, with local structural and kinematic complexities (Seranne 1992; Braathen et al. 2000; Osmundsen et al. 2003). The Devonian collapse was partly driven by major strike-slip shearing along reactivating Caledonian fault and shear zones (MTFC, HFZ, GGF, WBF, HBF) (Osmundsen and Andersen 1994; Dewey and Strachan 2003; Fossen 2010), during which the Devonian basins of southern Norway, East Greenland and Britain were formed (Seranne and Seguret 1987; Seguret et al. 1989; Fossen 1992) and lower crustal and high pressure metamorphic rocks were exhumed, as prominently displayed in the Western Gneiss Region in southern Norway (Andersen et al. 1991; Brueckner and van Roermund 2004; Hacker et al. 2010) and East Greenland (Hartz et al. 2001; Gilotti et al. 2014).
- In late Palaeozoic to Triassic(?) times, rifting still essentially reflected orogenic collapse, and NE-SW-oriented shallow, brittle crustal structures were reactivated that were favourably aligned with the dominant stress field.

- Beginning in the Triassic and particularly during the Jurassic, complex fragmentation of Pangea took place, with rifting including the dominant N-S trend of the Viking Graben and Halten Terrace. This extensional trend probably represents a linkage between Tethys and proto-Norwegian Sea. This period represents a long time interval (circa 100 million years) during which the dominant E-W stress field was highly oblique to the lithospheric-scale Caledonian orogenic structures, preventing full lithospheric rupture and breakup. This was probably a significant contributing factor in the anomalously long period between initial rifting and breakup in the North Atlantic (c. 350 million years).

- A major change in extension vector from E-W to NW-SE in the Early Cretaceous (Doré et al. 1999) resulted in favourable alignment with the pervasive mantle-lithospheric fabric and late-Caledonian shear zones. Major basins such as the Rockall, Faroe-Shetland, Møre, Vøring and Thetis basins, oblique to the earlier Jurassic N-S trend, had their principal expression at this time. These crust beneath the basins was hyperextended (Lundin and Doré 2011) but never achieved full oceanic status.

- Apparent cessation of extensional stress in the mid-Cretaceous (although with minor extension in some sub-basins) resulted in a significant time gap (60-70 million years) before final early Cenozoic breakup. This hiatus further contributed to the anomalously long time between initial rifting and breakup.
- Rupturing to form the North Atlantic was highly asymmetric in the south (*segment 2*) and oblique to the preceding basin trend in the north (*segment 1*) (Fig. 8). We suggest that the late Caledonian shear trend, deeply ingrained in the mantle, was now reactivated, but did not follow the axes of the previously formed basins that followed the shallower crustal trends and where cooling and strengthening may have occurred (e.g. Naliboff and Buiter 2015). The obliquity in *segment 1*, where the Thetis and Vøring basins may represent a single basin that has been diagonally bisected by breakup (Fig. 8), is an interesting issue; it is difficult to see why basin-parallel breakup akin to that of the Møre Basin did not occur. Cut-across of the basin by pre-breakup strike-slip, either newly formed or reactivating elements of the late Caledonian shear fabric, is one potential explanation for this geometry (Lundin & Doré, 2018; see also Section 7.2). Additionally, the oblique geometry of this segment may have been also controlled by pre-breakup magmatic intrusions.

One model for rift relocation suggests strengthening of the lithosphere by cooling after cessation of initial rifting (Van Wijk and Cloetingh 2002; Naliboff and Buiter 2015). This model might apply to the abandonment of the Møre and Rockall-Hatton shelves (Gernigon et al. this volume; Kimbell et al. 2017; Guan et al. 2019). However, this would not explain why the initial rifting stopped in the first place. Another mechanism is strain hardening in the vicinity of rift basins that may lead to development of a new rift offset from the early one (Kusznir and Park 1987; Sonder and England 1989; Newman and White 1997; Yamasaki and Stephenson 2009).

7.2 Magmatism, rifting and breakup

In the North Atlantic, variations in the magmatic budget during the onset of breakup were often controlled by complex interactions between the pre-existing lithosphere state (including discrete pre-existing structures, lithospheric thickness variations, thermal state and composition) and the timing and degree of decompression melting (Gernigon et al. this volume; Gouiza and Paton 2019).

As discussed earlier, the amount of pre- and syn-breakup decompression melting beneath continental margins is dependent on extension rate and pre-existing lithospheric rheology and composition (Buck 1991; Armitage et al. 2010; Huismans and Beaumont 2011; Petersen and Schiffer 2016), as well as on the geotherm (White and McKenzie 1989; Hill 1991). Petersen & Schiffer, (2016) suggest that a hot, weak crust over a relatively strong lithospheric mantle can produce wide, asymmetric, magma-rich margins. In contrast, a cold, strong crustal layer above a weaker mantle lithosphere may facilitate, magma-poor margins with abrupt necking zones (Petersen & Schiffer, 2016). As indicated in section 7.3, the abrupt margins observed in the NE Atlantic between Greenland and Norway may also related to exploitation of deep-seated and lithospheric-scale shear faults (Lundin & Doré, 2018).

Literature on the origin of magma-rich margins in the North Atlantic is dominated by the plume concept; in this hypothesis, the impingement of the Icelandic plume on the base of the lithosphere has variously been implicated in raised mantle temperatures, elevated margins, voluminous magmatism and break-up itself. A full description of this model is beyond the scope of this paper; it has been well-described in (for example) (White and McKenzie 1989; White 1992; Skogseid et al. 1992, 2000) while problems with the hypothesis have been highlighted by (for example) (Foulger 2002, 2010; Lundin and Doré 2005). Other ideas exist to explain the anomalous magmatism. These include a relationship to extension rate during breakup (Lundin et al. 2014) and the generation of small-scale edge-driven convection at abrupt steps in the lithosphere (Mutter et al. 1988; van Wijk et al. 2001).

Independent of the origin of anomalous magmatism in the North Atlantic, magmatic processes guided by pre-existing faults and shear zones may have influenced and governed final plate separation. Lithosphere softening associated with melts and lithospheric hardening associated with emplaced and cooled mafic rocks can occur at different depths and can accommodate and localise strain. Magma-supported lithospheric breakup may occur at far lower differential stress levels than those needed for lithosphere breakup via brittle faulting (Buck and Karner 2004).

Most of the magmas feeding plateau basalts and SDRs in the North Atlantic are associated with large Palaeocene to Eocene igneous centres (Callot et al. 2001; Callot and Geoffroy 2004; Geoffroy et al. 2007). These may have been related to small-scale convection cells that initiated at the base of the lithosphere and grew upward by thermal erosion, feeding these localised igneous centres and creating “soft spots” in the lithosphere (Geoffroy et al. 2007; Gac and Geoffroy 2009). These small-scale convection cells appear to correlate with areas of high mantle heat flow suggesting a relationship with lithospheric thickness variations (Geoffroy et al., 2007). The pattern and development of such instabilities, marking the potential locus of the future breakup axis, could thus reflect the pre-existing thermal, compositional and structural configuration of the lithosphere.

Cenozoic breakup of the NE-Atlantic did not occur within previously (Late Jurassic-Early Cretaceous) tectonically thinned lithosphere, such as the Hatton-Rockall shelf. These areas thermally re-equilibrated and strengthened after early rifting events (Guan et al., 2019; Gernigon et al., 2019). On the Hatton-Rockall shelf, the distance between the Palaeogene igneous centres is approximately 100 km – about twice as large as along the continental margins of East Greenland and Hatton-Rockall (Geoffroy et al. 2007; Horni et al. 2017).

In the British Tertiary Igneous Province an abnormally dense spacing of igneous centres is observed (Doubre and Geoffroy 2003). This dense pattern may indicate that the pattern of Palaeogene small-scale convection or loci of mantle diapirism interacted with compartmentalised lithospheric blocks and terranes originating with Caledonian and post-Caledonian shear motions. The developing igneous centres then exploited lithospheric structural and compositional heterogeneities.

Edge convection along the eastern border of the Greenland craton (King and Anderson 1998) may have increased magmatic production rates and reduced the spacing between small-scale convection cells, forming igneous centres and weakening and thinning the lithosphere. This model could be one explanation for the development of the breakup axis oblique to most failed rift systems in the NE Atlantic (Gernigon et al., 2019), along with an oblique inherited lithospheric mantle fabric and lithospheric perforation by pre-syn-breakup strike-slip motion (see next section), or a combination of these end-member mechanisms (Fig. 12).

7.3 Role of strike-slip faults

Lundin & Doré (2018) recently proposed that the instigation of oceanic spreading in the North Atlantic-Arctic region was facilitated by the development of transform faults. Such faults are strike-slip faults that segment plates or form plate boundaries, juxtaposing oceanic and continental crust. According to Lundin and Doré (2018), some of these faults were inherited structures while others formed first during the breakup process. The basis of this model is that a) pre-existing lines of lithospheric-scale strike-slip faults are zones of weakness, which can be separated by orthogonal forces without initial stretching, and b), that oblique slip more easily facilitates breakup (Brune et al. 2012a). Thus, such zones should fail first (Brune et al. 2018).

A clear example of a margin influenced by transform faulting is found on the SW Barents Sea margin, which opened along the De Geer transform fault. This fault was probably instigated during late-stage sinistral movements along the Caledonian Orogen in the Late Devonian (Harland 1969), but its role in oceanic development did not start until the Eocene opening of the NE Atlantic, when it enabled Greenland to be translated dextrally past Eurasia. In the earliest Oligocene, the De Geer transform fault opened obliquely as a zone of deformation that developed into an oblique transform margin, along which the Knipovich Ridge ultimately formed between Eurasia and Greenland (Faleide et al. 2008). The De Geer transform fault and rigid crustal/lithospheric blocks in the SW Barents Sea may have acted as a barrier to the straight propagation of the North Atlantic. A further example is found in the southernmost CNAR, where the Aptian opening of the Bay of Biscay utilised the North Pyrenean Fault, an established Variscan transform fault (Vissers et al. 2016).

Margins influenced by strike-slip faulting may be characterised by the reduced role of extension prior to breakup as well as the potential for the breakup axis to be oblique to earlier rifting. The NE Atlantic, as shown earlier, is an example where older (Mesozoic) rift systems are cut obliquely by the axis of Early Eocene breakup, resulting in highly asymmetric conjugate margins (Lundin et al. 2013).

Based on these observations, Lundin & Doré (2018) suggested that the old rift systems were bisected by transform faulting, which facilitated orthogonal opening in the Early Eocene. Kinematic evidence along the line of breakup provides some support for this hypothesis. Indications of dextral motion are provided by the Hel Graben in the northern Vøring Basin. The Hel Graben would be located at a right-stepping releasing bend in the proposed pre-NE Atlantic shear, offset by the Surt Lineament (Blystad 1995; Brekke 2000). Within the graben are a series of E-W trending normal faults (Ren et al. 2003), consistent with cross-basin fault systems in a dextral pull-apart basin (Dooley and McClay 1997).

The strike slip deformation and associated breakup may have acted rapidly (on geological scale) along long segments of the North Atlantic rift. Such a model is consistent with the rapid lithospheric relaxation observed in the shift of depocentres in the Danish Basin (Nielsen et al. 2007). Other authors prefer a rift propagation model characterised by highly diachronous and fragmented breakup (Gernigon et al. this volume). In such a model, the North Atlantic rift would have propagated along the weakest path for the background stresses and this may have been governed by basement inheritance or pre-syn-rift magmatism that has perforated the lithosphere. These two models may not necessarily be incompatible: A rift propagation model could have had a strong component of strike-slip deformation, or strike-slip may have acted only on certain segments rather than the whole length of the NE Atlantic margins.

Whether or not one accepts the evidence for strike-slip motion immediately preceding breakup in the NE Atlantic, a further key observation is that of the close correlation of the breakup axis and NE-SW-trending late Caledonian shears, such as the MTFC and GGF. This suggests that a more ancient and deep-seated strike-slip trend was implicated in the eventual line of opening.

Both, the Labrador Sea and Baffin Bay appear to cut through cratonic elements and across Proterozoic orogenic belts (Section 1.3) (St-Onge et al. 2009). As it is unusual that such features would form in a mid-cratonic setting, the possibility exists that they were facilitated by transform faults. Indirect support for such a fault in the proto-Labrador Sea is provided from the c. 1000 km long fault system in Hudson Strait and Foxe Channel, the projected continuation of the Canadian Labrador Sea margin. The fault system has been interpreted as an abandoned rift tip to the Labrador Sea rift (Pinet et al. 2013), but vertical offsets are small compared with the length of the fault system, and terminate northwards in a horse-tail geometry.

The fault system is marked by shallow rhomboid basins, suggesting sinistral movement. It may, therefore, have originated as a sinistral transform that experienced minor extensional overprint during early opening of the Labrador Sea. When the Ungava Transform linked the Labrador Sea and Baffin Bay in the Palaeogene (Funck et al. 2012), the Hudson Strait-Foxe Basin fault system was abandoned. However, the Labrador Sea transform fault might have had a much older origin. Whereas there is no evidence of a suture beneath the Labrador Sea, there is abundant evidence on the Labrador margin of sinistral shear sub-parallel to the ocean, which van Gool et al. (2002) and St-Onge et al. (2009) relate to the Palaeoproterozoic (c. 1.8 Ga) indentation of the North Atlantic Craton into northern Greenland. Thus, it is possible that these older shears were exploited during development of the Labrador Sea in a similar way as suggested for the NE Atlantic.

7.4 Microcontinent formation

A complication in North Atlantic breakup was formation of the JMMC (Gaina et al. 2009; Blischke et al. 2011, 2017; Schiffer et al. 2015b, 2018) and other smaller continental fragments (Døssing et al. 2008; Péron-Pinvidic and Manatschal 2010; Nemčok et al. 2016). The central segment of the NE Atlantic, between the GIFR and the Jan Mayen Fracture zone (Figure 1,2), underwent breakup and initially fast spreading along the Aegir Ridge in the early Eocene (~55 Ma) that then slowed down in the mid-Eocene (~47 Ma) (Gernigon et al. this volume, 2015). In the mid-Cenozoic, the JMMC began to separate from East Greenland's Liverpool Land margin along the new Kolbeinsey Ridge, and the Aegir Ridge became extinct, between ~28 Ma and ~21 Ma (Nemčok et al. 2016; Lundin and Doré 2018), when the JMMC separated from the East Greenland margin. The two mid-oceanic ridges were, therefore, simultaneously active for possibly up to 10 Ma, but probably longer (Doré et al. 2008; Gernigon et al. 2012, 2015; Peron-Pinvidic et al. 2012; Ellis and Stoker 2014).

The JMMC consists of Cenozoic igneous rocks and older thinned and intruded continental crust (Kuvaas and Kodaira 1997; Breivik et al. 2012; Blischke et al. 2017). Breakup on the eastern side of the JMMC was magmatic, forming subaerial seaward-dipping reflectors (SDRs) (Planke and Alvestad 1999) underlain by HVLCBs (Breivik et al. 2012). SDRs are not observed along the western margin of the JMMC (Kodaira et al. 1998), nor are they reported from the conjugate Liverpool Land margin (Horni et al. 2017). Wide-angle seismic data suggest that the northern part of the JMMC is underlain by "Icelandic-type" crust (Kandilarov et al. 2015). The central Jan Mayen Ridge comprises ~15 km thick continental crust (Kodaira et al. 1998; Breivik et al. 2012) with no evidence of HVLC (Kodaira et al. 1998; Mjelde et al. 2007a), but HVLC is observed in the transition zone to the Iceland plateau (Gernigon et al. this volume; Brandsdóttir et al. 2015).

The mechanisms responsible for breakoff of the JMMC are poorly understood. Plume impact has been suggested (Müller et al. 2001; Mittelstaedt et al. 2008; Howell et al. 2014). In such a scenario a plume heats and weakens the lithosphere to cause renewed continental breakup and a ridge jump. However, breakup-related volcanism, such as SDRs, are absent between Greenland and the JMMC (Kodaira et al. 1998; Horni et al. 2017), at odds with expectations of this model.

Also mechanical explanations for the formation of microcontinents and rifted continental blocks via detachment along pre-existing lithospheric weaknesses have been suggested (Nemčok et al. 2016; Molnar et al. 2018; Schiffer et al. 2018). Recent analogue modelling illustrates how microplates may be formed by propagating rifts that form new oceans. Microcontinents may separate from the continental margin with rotational motion during the latest breakup stages, such that location and shape of fragmentation is controlled by lithospheric weaknesses (Molnar et al. 2018). Analogue modelling produces rotating microplates "trapped" between overlapping spreading centres (Katz et al. 2005). Geoffroy et al. (2015) suggested microcontinent formation though large-scale, symmetric and continental-vergent detachment faulting – the so-called "C-Block". Other models involve the separation of continental lithosphere along overlapping spreading centres (Auzende et al. 1980; Ellis and Stoker 2014). Foulger et al. (this volume) suggest that extension in the southern JMMC was initially diffuse, and westward migration of the axes of extension on the GIFR induced extension in the JMMC, focusing on its most westerly axis to form the proto-Kolbeinsey Ridge. This resulted in breakoff of the JMMC from Greenland.

Several authors have linked separation of the JMMC to mid-Eocene plate-kinematic reorganisations in the North Atlantic (Gaina et al. 2009). For instance, Schiffer et al. (2018) proposed a model linking the formation of the JMMC to global plate tectonic reconfigurations to rejuvenation of old and pre-existing lower-crustal/upper mantle orogenic fabric (Schiffer et al. 2015b; Petersen and Schiffer 2016). In this model, the apparent rotation of North Atlantic spreading from NW-SE to W-E put the NW-SE oriented accommodation zone between the Aegir and Reykjanes Ridges – the proto-GIFR – under transpression. This "locked" the right-lateral deformation along the proto-GIFR and forced extension to divert to a more favourable position/path – possibly pre-existing Caledonian weak zones along the East Greenland margin.

7.5 The Wilson Cycle revisited

The breakup of Pangaea to form the NE Atlantic was a protracted and piecemeal process with many local complexities (Gernigon et al. this volume; Peace et al. this volume; Roberts et al. 1999; Ady and Whittaker 2018). The original Wilson cycle theory does not conclusively explain how and why the opening of the North Atlantic occurred and why its manifestation varies across the CNAR. During the past 50 years, more data have been acquired, and new theories proposed, but the mechanisms driving the Wilson Cycle are still a matter of debate. The simplest explanation for breakup along former orogens is that mountain ranges are usually the weakest zones in supercontinents and hence regions where deformation is expected to concentrate.

We show that most rifts are associated with former collision zones, implying that structural inhomogeneity may be preserved long term. However, the North Atlantic did not necessarily focus exactly along suture zones, but in some places broke through regions of apparently previously undisturbed cratonic lithosphere such as SE Greenland (Buiter and Torsvik 2014). The Labrador Sea and Baffin Bay broke through pre-existing cratons (the Archean North Atlantic and Rae cratons) and almost orthogonally across Precambrian orogenic belts (the Meosoproterozoic Grenville and Makkovik-Ketilidian orogens, and the Paleoproterozoic Nagssugtoqidian orogen) (Buchan et al. 2000; St-Onge et al. 2009; Peace et al. 2017).

These events may have been enabled by the development of transform faults that can also nucleate at a distance from old suture zones (Lundin & Doré, 2018), or by the regional rift, magmatic and thermal history that modified lithospheric strength distribution and guided lithospheric breakup away from suture zones – for example, the pronounced N-S rift fabric that developed in the Jurassic, oblique to the main Caledonian trend. Alternatively, they may have occurred simply for kinematic compatibility reasons, for example if the pathway through a craton was the shortest.

The Wilson Cycle is one of the most crucial and basic concepts regarding inheritance in a plate tectonic framework. However, the Wilson Cycle concept only addresses large-scale, first-order events and is insufficient to explain all the complexities of developing oceans and continental margins. (Super)continents do not simply re-open along the surface traces of suture zones.

Structures and fabric are 3D entities and dominant weaknesses are not at the surface. Inheritance at all scales is important in explaining rejuvenation at regional and global scales. Structures can be preserved over billions of years and may still impose an inheritance control. Intraplate deformation and "non-rigid" plates and magmatism must also be incorporated in inheritance models and plate tectonic theory.

8 Conclusions

1. The CNAR is the type example of the Wilson Cycle concept and, in general, reopened elements of the Caledonian fold belt. However, evidence from the CNAR and elsewhere clearly demonstrates that the Wilson Cycle only partially accounts for the observations. Where breakup does occur along an older orogenic belt, it does not simply re-open the older suture, and intact cratons may fragment.
2. Rift evolution and opening of the CNAR is to varying degrees linked to structural inheritance at lithospheric, continental, basin, fault and micro-scales. However, other factors were influential, such as changing stress directions imposed by plate boundary effects and supercontinent (Pangaea) breakup and magmatism.
3. Caledonian rejuvenation of anisotropies imparted at different depths (crust vs. mantle) and different stages of the orogenic evolution (collision vs. late Caledonian transpression-transtension) played the major role in regional rift evolution and breakup of the NE Atlantic. Precambrian structures (e.g., the Nagssugtoquidian suture, Bothnia-Senja Fault Zone) may have also controlled the margin segmentation of the NE Atlantic on the largest scale.
4. Many, but not all, of the rift systems that preceded the CNAR followed major orogenic trends expressed at the surface. However, final breakup seems to have followed the late Caledonian strike-slip shear fabric exemplified by the MTFC. In the NE Atlantic (*Segment 1*), breakup cut obliquely across the preceding rifts. We suggest that breakup occurred when stress directions became favourable to exploit a deeper and more pervasive mantle fabric, probably related to lithospheric-scale shear zones that developed in the late Caledonian. The radical cut-across of a Cretaceous basin by the breakup line in Segment 1, defining the Thetis and Vøring margins, may represent reactivation of this trend by pre-breakup strike-slip and/or was guided by pre-breakup magmatic intrusions weakening the crust.
5. Breakup between SE Greenland and the Rockall-Hatton margin does not appear to fit the classic Wilson-Cycle model. This may have been be related to the pre-breakup rift history of the Hatton-Rockall shelf which thinned crust but created an overall stronger lithospheric column. Breakup occurred where the crust was thicker above weaker lithosphere, and may have been assisted by strike-slip/transform motion. This region then formed the southern CNAR link, which was offset from the Northern CNAR breakup axis between the Central Atlantic and the Aegir Ridge
6. The extremely long interval between initial post-orogenic rifting and final plate separation – some 350 million years – is highly anomalous and an order of magnitude greater than that of most other oceans – for example the Central and South Atlantic. It was probably a result of radical changes in the regional stress field – for example in the Triassic-Jurassic interval, and significant hiatuses in extension (for example in the mid-Late Cretaceous), which allowed the pre-existing rifts to cool and strengthen.

7. The “magma-poor” nature of the western margin of the JMMC and adjacent ocean favours mechanical models for microcontinent formation, rather than weakening by a thermal anomaly.
8. High velocity lower crustal bodies beneath the basins flanking the North Atlantic have been variously attributed to syn-rift serpentinised mantle, syn-rift mafic and ultramafic rocks and pre-existing metamorphic and metasomatised rocks such as granulites and eclogites. Some of these interpretations and the question of whether these HVLCBs have a pre-or syn-rift origin remain controversial. Some HVLCBs formed before the onset of rifting and breakup, forming rheological inhomogeneities in the lithosphere that can control the location, deformation and type of breakup and continental margins.
9. The temperature- and composition-dependent strength profile of a lithospheric column, which controls crust vs. mantle thinning (among other properties), but also the extension rate, structural and rift obliquity determine whether a wide, asymmetric, diffuse rift zone develops or whether sharply localised rifts with narrow necking zones develop.
10. The GIFR may contain a significant component of continental lithosphere. It may owe its existence to diffuse extension of a zone of Precambrian terranes and Caledonian fossil structures running parallel to the direction of extension between central East Greenland and northern Scotland.
11. At the basin-scale, the CNAR displays a wide variety of structural inheritance effects including brittle reactivation of basement faults and fabrics creating syn-rift faults that are complex in terms of kinematics or growth history. Oblique extension on pre-existing orogenic, post-orogenic or early rift structures may partition deformation into strike-slip and dip-slip components at different scales. Complex fault displacement patterns are produced by multiphase rifting where later extension collinear with, or rotated relative, to earlier phases by local stresses. Basin inversion structures by definition are inherited features where intense localisation of contractional deformation is guided by pre-existing extensional features. Igneous activity and other mobilised materials such as salt interact with local stress fields to partition deformation within basins both spatially and temporally.

Acknowledgements

Christian Schiffer’s Carlsberg Foundation postdoctoral fellowship was held at Durham University, where the series of North Atlantic workshops that led to this special issue were planned and held. Thomas Phillip’s postdoctoral fellowship at Durham University is funded by the Leverhulme Foundation. M. Stoker gratefully acknowledges the award of Visiting Research Fellow at the Australian School of Petroleum. The authors thank all participants of the North Atlantic workshop series. Four anonymous reviewers and the handling editor are thanked for constructive and useful comments, and John Kipps (Equinor) for help with some of the figures.

9 Figures

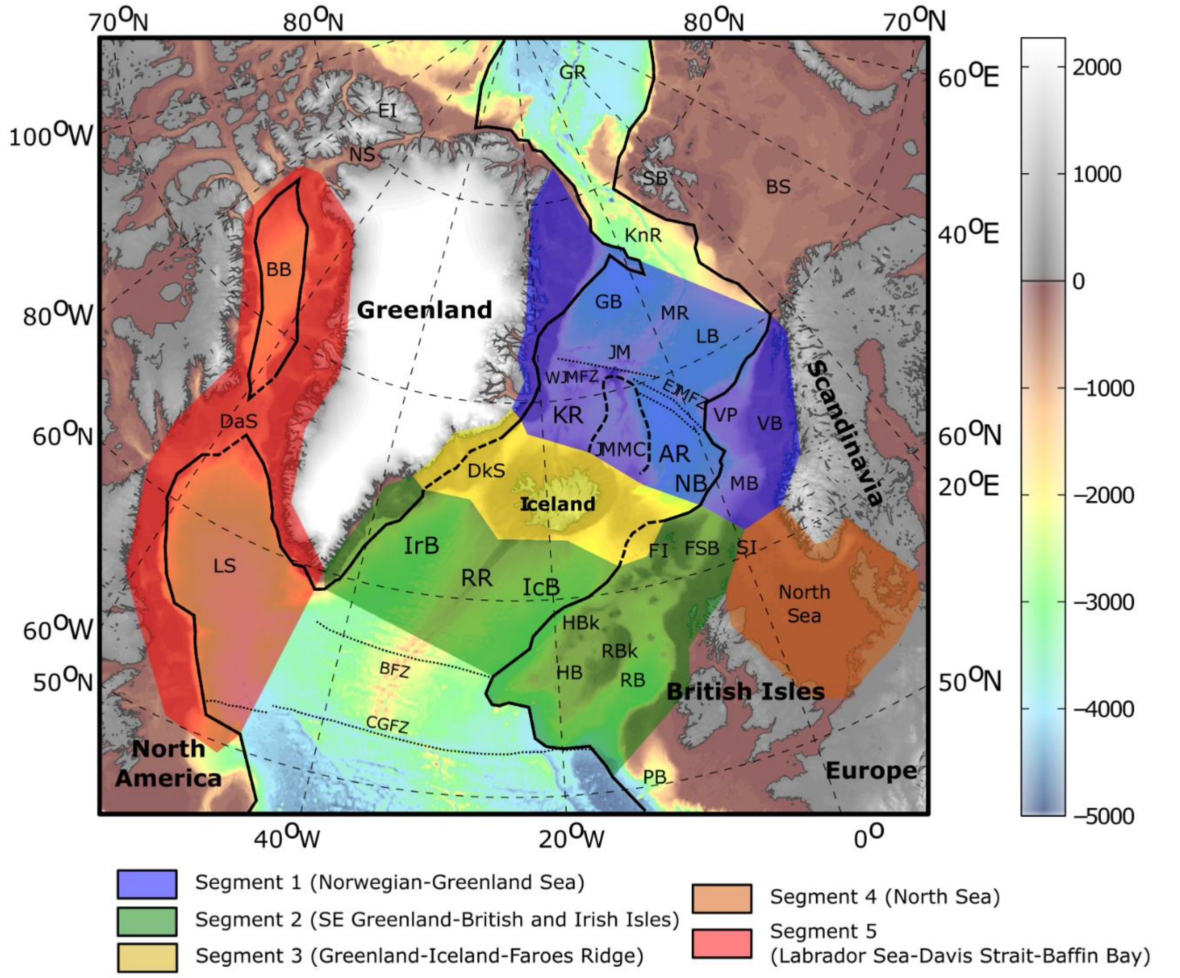

Figure 1: Physiographic map of the present-day Circum-North Atlantic region showing geographic names and places, as well as the five segments discussed in the paper. Abbreviations: AR = Aegir Ridge, BB = Baffin Bay, BFZ = Night Fracture Zone, BS = Barents Sea, CGFZ = Charlie Gibbs Fracture Zone, DaS = Davis Strait, DkS = Denmark Strait, EI = Ellesmere Island, EJMFZ = East Jan Mayen Fracture Zone, FI = Faroe Islands, FSB = Faroe-Shetland Basin, GB = Greenland Basin, GR = Gakkel Ridge, HB = Hatton Basin, HBk = Hatton Bank, IcB = Iceland Basin, IrB = Irminger Basin, JM = Jan Mayen Island, JMMC = Jan Mayen Microplate Complex, KnR = Knipovich Ridge, KR = Kolbeinsey Ridge, LB = Lofoten Basin, LS = Labrador Sea, MB = Møre Basin, MR = Mohn's Ridge, NB = Norway Basin, NS = Nares Strait, PB = Porcupine Basin, RB = Rockall Basin, RBk = Rockall Bank, RR = Reykjanes Ridge, SB = Svalbard, SI = Shetland Islands, VB = Vøring Basin, WJMFZ = West Jan Mayen Fracture Zone. Solid black lines are an interpretation of the continent-ocean transition. Stippled black lines indicate uncertain locations of the continent-ocean transition.

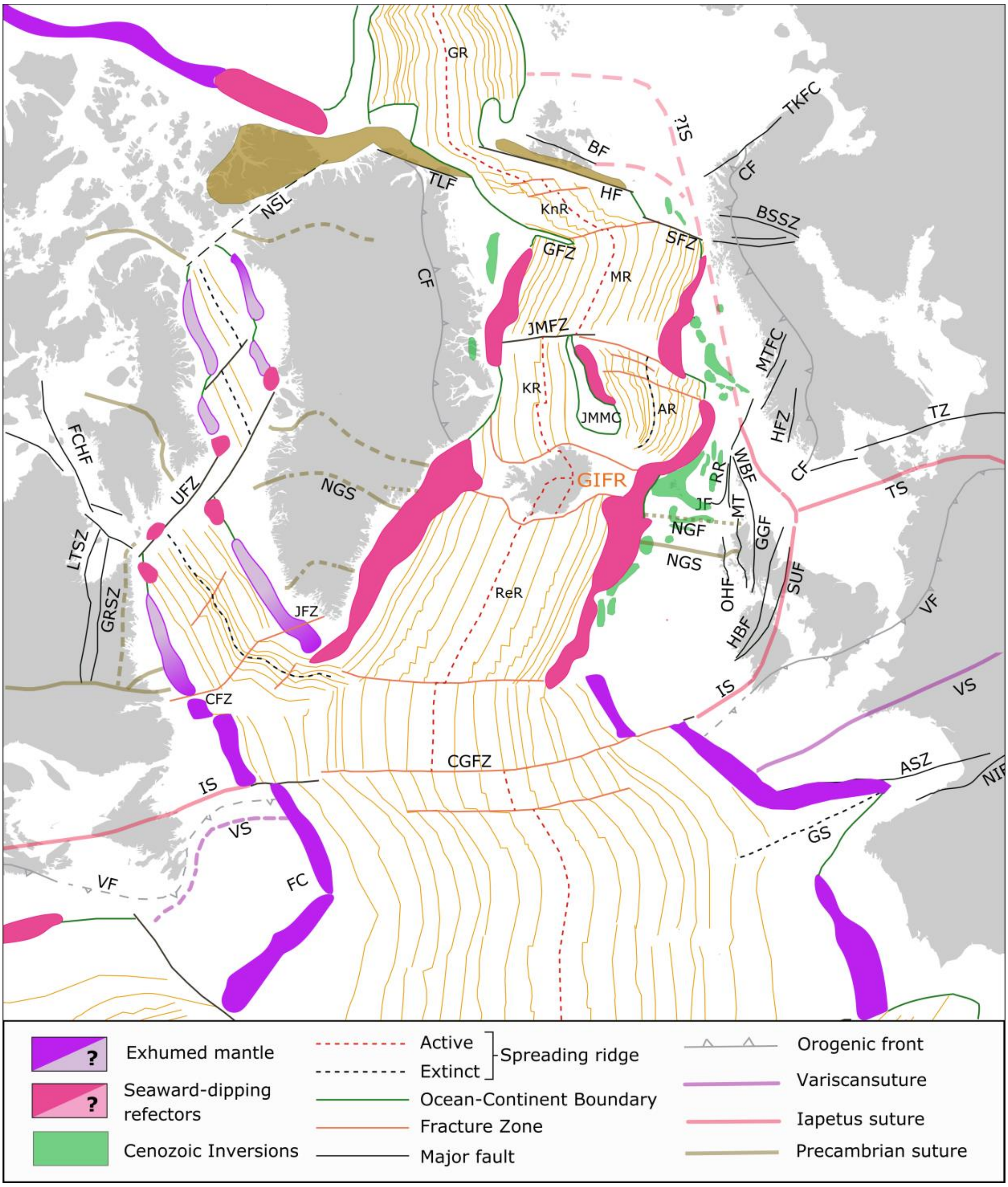

Figure 2: Structural map of the present day Circum-North Atlantic region, showing oceanic structure, marginal character and major onshore basement lineaments. Abbreviations: AR= Aegir Ridge, ASZ = Armorican Shear Zone, BF = Billefjorden Fault, BSSZ = Bothnian-Senja Shear Zone, CF = Caledonian front, CFZ=Cartwright Fracture Zone, CGFZ = Charlie Gibbs Fracture Zone, FC = Flemish Cap, FCHF = Foxe Channel – Hudson Bay fault system, GFZ = Greenland Fracture Zone, GGF = Great Glen Fault, GIFR – Greenland Iceland Faroes Ridge GRSZ = George River Shear Zone, GS = Goban Spur, HBF = Highland Boundary Fault, HF = Hornsund Fault, HFZ = Hardangerfjord Fault Zone, IS = Iapetus Suture, JF = Judd Fault, JFZ=Julianehaab Fracture Zone, JMFZ = Jan Mayen Fracture Zone, JMMC = Jan Mayen Microplate complex, KnR= Knipovich Ridge, KR= Kolbeinsey Ridge, LTSZ = Lac Tudor Shear Zone, mR= Mohn's Ridge, MT = Moine Thrust,

MTFC = Møre-Trøndelag Fault Complex, NGF = Nagssugtoquidian Front; NGS = Nagssugtoquidian Suture, NIF = North Iberian Fault, NSL = Nares Strait lineament, OHF = Outer Hebrides Fault, ReR= Reykjanes Ridge, RR = Rona Ridge, SFZ = Senja Fracture Zone, SUF = Southern Uplands Fault, TKFC = Trollfjord-Komagelv Fault Complex, TLF = Trolle Land Fault, TZ = Tornquist Zone, TS = Thor Suture, UFZ = Ungava Fault Zone, VF = Variscan Front, VS = Variscan Suture, WBF = Walls Boundary Fault.

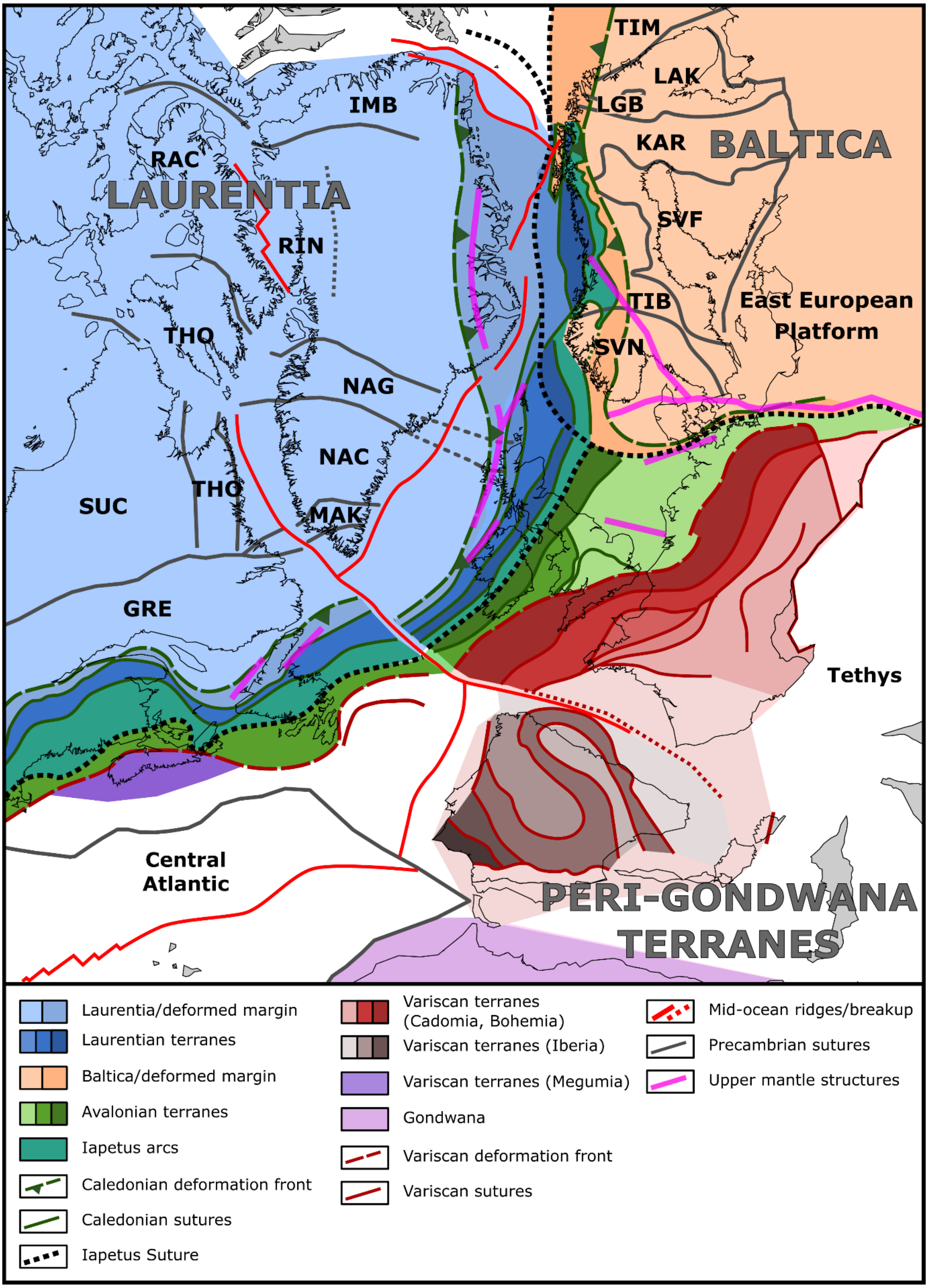

Figure 3: Terrane and structure map of the Circum-North Atlantic region at 145 Ma showing continents, terranes, suture zones and upper mantle structures. Caledonian and Variscan terranes and structures are generally closely aligned with the breakup axes In the NE Atlantic, breakup occurred west of the Iapetus Suture; In contrast, breakup occurred to the east of the Iapetus Suture in the northern Central Atlantic. Precambrian terranes are

generally perpendicularly aligned, but have therefore probably affected margin segmentation and the formation of transform zones and faults during rifting. The NW Atlantic is a region where an ocean opened (Labrador Sea and Baffin Bay) not following any known Phanerozoic structures or terranes and cross-cutting Precambrian lineaments. However, the Rinkian Orogen may have played a role during the formation of Baffin Bay. GRE – Grenvillian Orogen, IMB – Inglefield Mobile Belt, KAR – Karelian, LGB – Lapland Granulite Belt, MAK – Makkovik-Ketilidian Orogen, NAC – North Atlantic Craton, NAG – Nagssugtoqidian Orogen, RAC – Rae Craton, RIN – Rinkian Orogen, SUC – Superior Craton, SVF – Svecofennian, SVN – Sveconorwegian Orogen, THO – Trans-Hudson Orogen, TIB – Transscandinavian Igenous Belt, TIM – Timanian Orogen

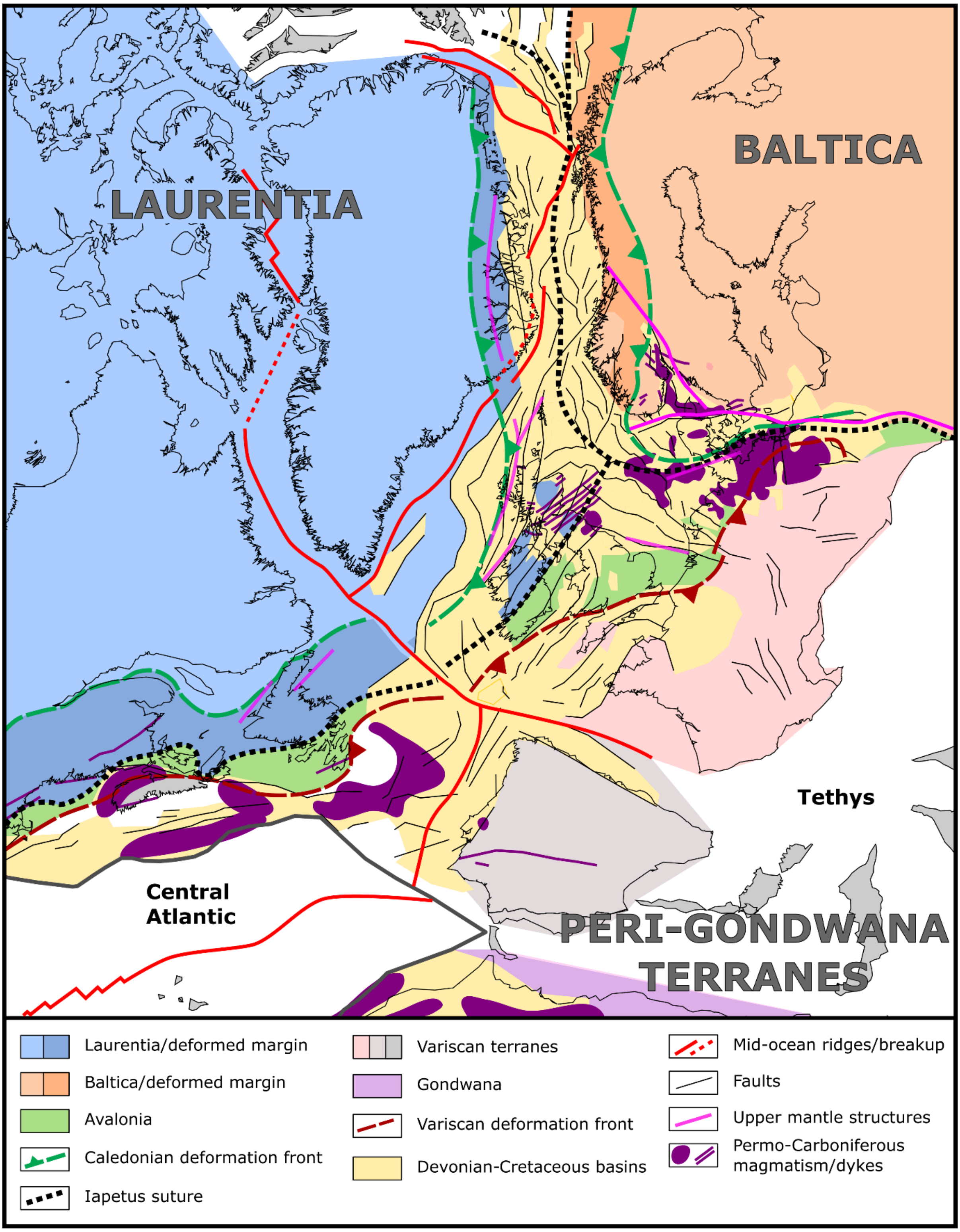

Figure 4: Basin and structure map of the Circum-North Atlantic region at 145 Ma showing continents, basins, faults, upper mantle structures and Permo-Carboniferous magmatism. Almost all of the early, Devonian-Jurassic basins illustrated in this figure have formed within the Caledonian or Variscan orogens (within their deformation fronts) and oblique to the axis of breakup.

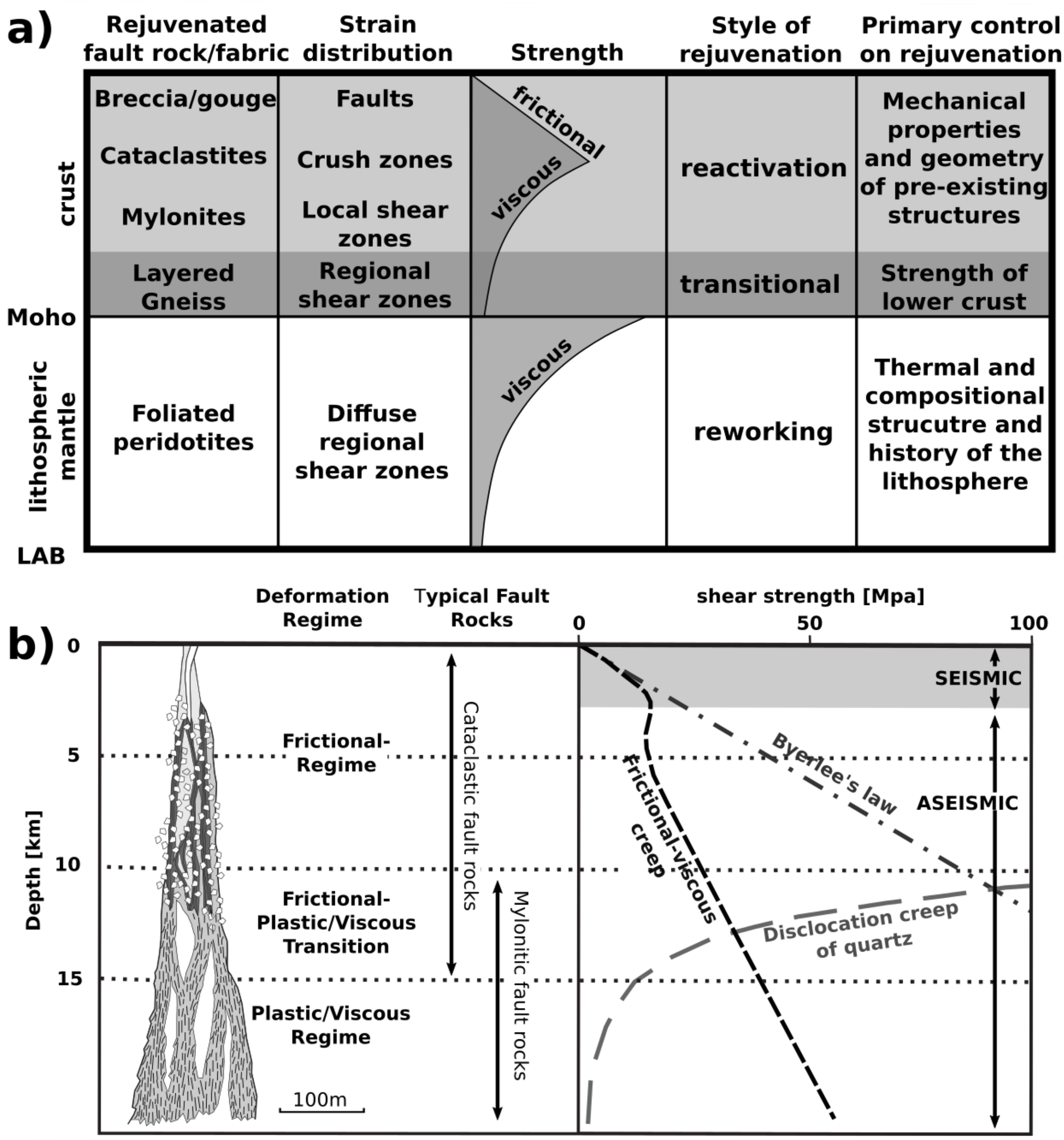

Figure 5: (a) Schematic diagram showing typical fault rocks/fabrics, the strain distribution, strength, tectonic style and primary rheological controls during rejuvenation at different depth through the lithosphere (Holdsworth et al. 2001; Jefferies et al. 2006)***. The regimes of reactivation and reworking are separated by a gradual transition somewhere in the lower crust. Note that the strength profile is for a simplified and averaged continental lithosphere. Any tectonic processes (orogenesis, delamination, rifting) and compositional heterogeneity will perturb the strength profile of the lithosphere. (b) Schematic diagrams illustrating variations in shear deformation, deformation regime and typical fault rocks with depth in a crustal profile, (left) and a crustal strength profile with different representative rheologies (right)*** (Holdsworth et al. 2001; Alsop and Holdsworth 2004; Jefferies et al. 2006; Imber et al. 2008)***.***

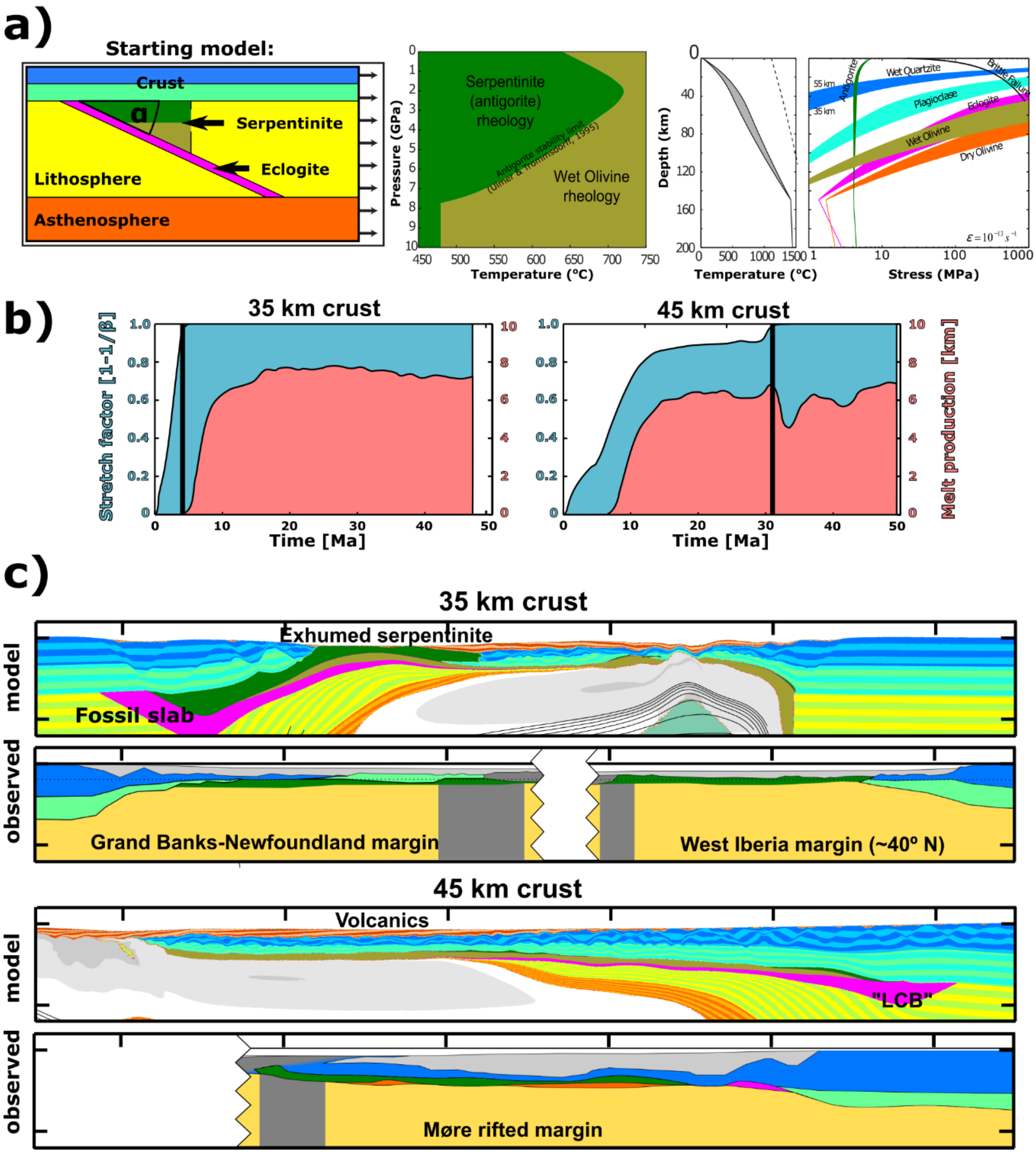

Figure 6: 2D numerical modelling setup and results modified from Petersen & Schiffer (2016) illustrating the effect of crustal thickness and a preserved subduction zone complex on rifting and passive margin formation. The crust vs. mantle (depth-dependent) thinning is in agreement with many other studies (e.g. Buck, 1991; Huismans and Beaumont, 2011) (a) Starting model setup with crust (upper and lower), lithospheric mantle with discrete heterogeneities (eclogite, serpentinite and hydrated peridotite) on top of the asthenospheric mantle (upper panel). The binary phase diagram for antigorite/serpentinite stability is shown (lower left panel), the different tested initial geotherms for a range of crustal thicknesses (35-55 km) (lower middle panel) as well as the resulting strength profiles for the involved lithologies (wet quartz, plagioclase, dry and wet olivine, and antigorite/serpentinite) (lower right panel). (b) Modelling evolution in terms of relative crustal thinning (blue, stretch factor = 1-1/β; 0 is no thinning, 1 represent separation of the continental crust – marked by the vertical black line) and melt production (red, in terms of equivalent thickness of flood

basalts). It can be noticed that melt production first starts with separation (breakup) of the continental crust, therefore, no flood basalts cover the continental crust. For a thick crust (45 km) it can be observed that melt production starts several tens of Ma before crustal separation (breakup), therefore intruding and extruding magmatic products into and on top of continental crust. (c) Models and possible analogues in the North Atlantic passive margins. The thin crust template is able to explain many first-order observations in magma-poor margins, such as the Iberia-Newfoundland conjugate passive margins. Here, the surrounding continents have thinner crust (~35 km). The margins have sharp, abrupt necking zones, thin continental slivers, separated from the margin lie within exhumed and hydrated mantle lithosphere covering the ocean floor and few (pre-breakup) magmatic products are observed on the continental margin. Volcanic products cover hundreds of km of the preserved hyperextended continental crust. The conjugate (not shown here) would be – in contrast – very narrow forming a highly asymmetric conjugate margin pair. This "magma-rich" margin shows evidence of high velocity lower crust that – in this case – is derived from the deformed and re-emplaced lithospheric heterogeneities. This template shows many similar first order features with many observed magma-rich margins, for example the Møre margin.

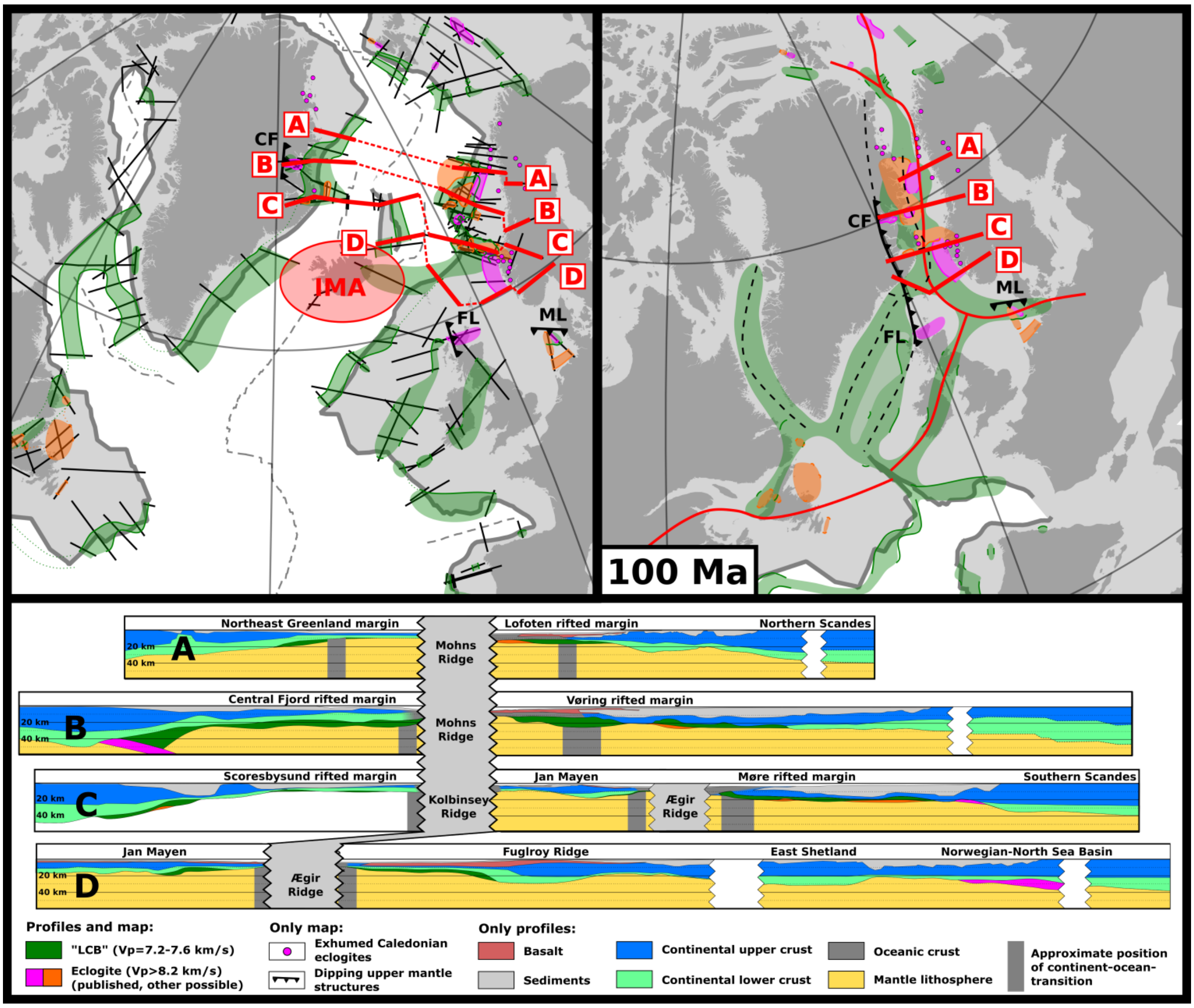

Figure 7: Top left: Map of the high-velocity lower crustal bodies in the Circum-North Atlantic region. Green colours illustrate HVLCBs in the P-wave velocity range 7.2-7.6 km/s

and magenta shows occurrences of ultra-HVLCBs with P-wave velocity larger than 8.2 km/s that is indicative of eclogite. Thick black lines show dipping upper mantle structures and triangles the dip direction: CF, Central Fjord structure; FL, Flannan structure; ML, Mona Lisa Caledonian suture. Orange indicates other possible occurrences of the Vp>8.2 km/s ultra-HVLCBs. Four transects are defined through the NE Atlantic based on seismic lines (thin black lines) illustrates the position of the transects (red) and wide-angle seismic lines (black lines) in the North Atlantic. Top right: Same map in a 100 Ma pre-breakup reconstruction showing how well the interpreted eclogite bodies coincide with the location of the Iapetus Suture (thin red line). Lower panel shows the four transects (A-D) at present day from wide-angle seismic lines (Theilen and Meissner 1979; Goldschmidt-Rokita et al. 1994; Weigel et al. 1995; Mandler and Jokat 1998; Christiansson et al. 2000; Tsikalas et al. 2005; Mjelde et al. 2009, 2013; Roberts et al. 2009; Voss et al. 2009; Breivik et al. 2012; Maupin et al. 2013; Schiffer et al. 2015a)**.**

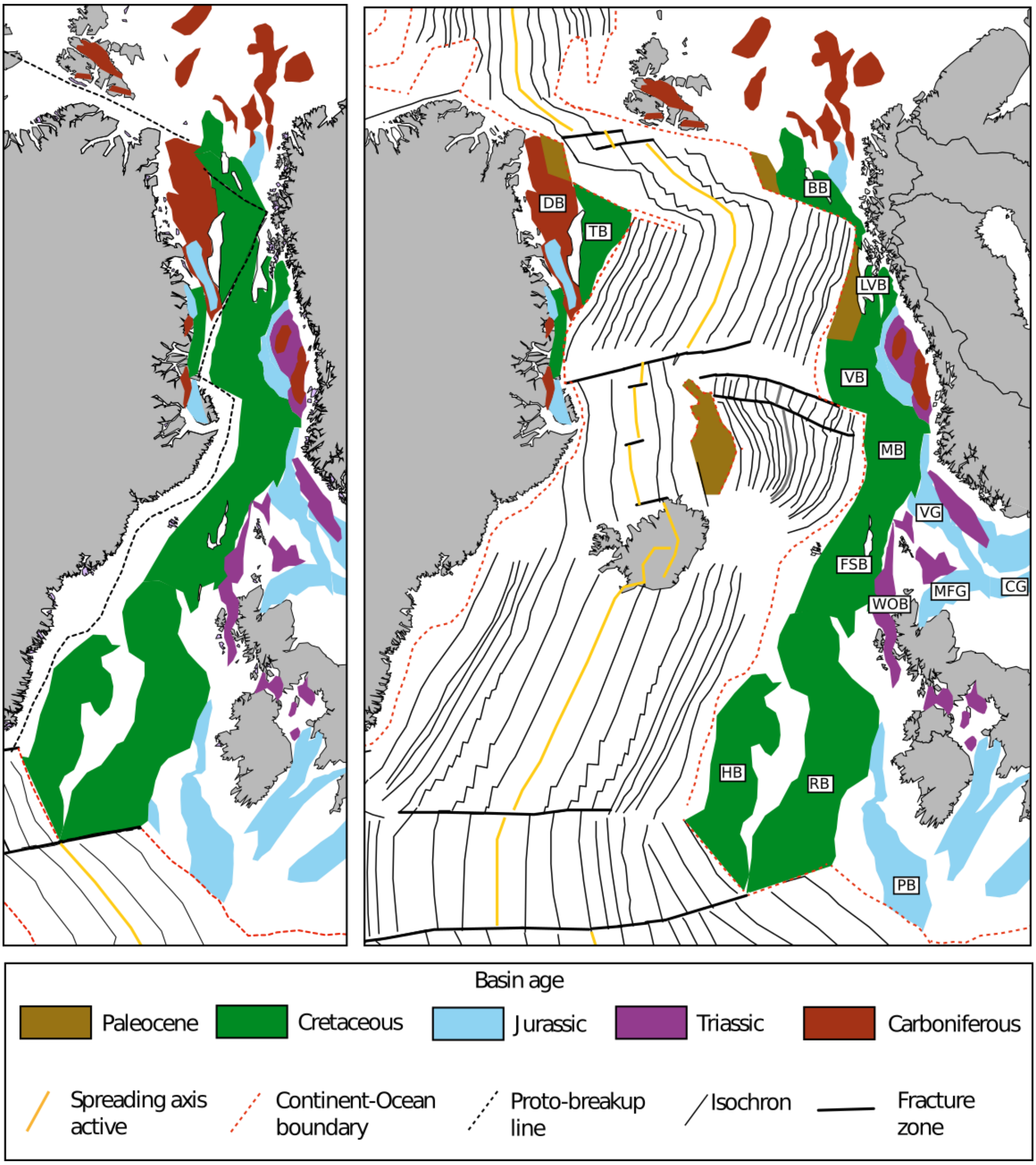

Figure 8: Basin age map of the North Atlantic, shown before breakup at 53 Ma (left), and present day (right). Basins are coloured according to the age of the crustal extension that mainly created the basin. The general asymmetry of the breakup (much of pre-existing basin system left on the European margin) and the obliquity of the breakup in the NE are clearly shown. Abbreviations: BB, Bjørnøya Basin; CG, Central Graben; DB, Danmarkshavn Basin; FSB, Faroe-Shetland Basin; HB, Hatton Basin; LVB, Lofoten-Vestrålen Basin; MB, Møre Basin; MFG, Moray Firth Graben; PB, Porcupine Basin; RB, Rockall Basin; TB, Thetis Basin; VB, Vøring Basin; VG, Viking Graben; WB=Wandel Sea Basin, WOB, West Orkney Basin.

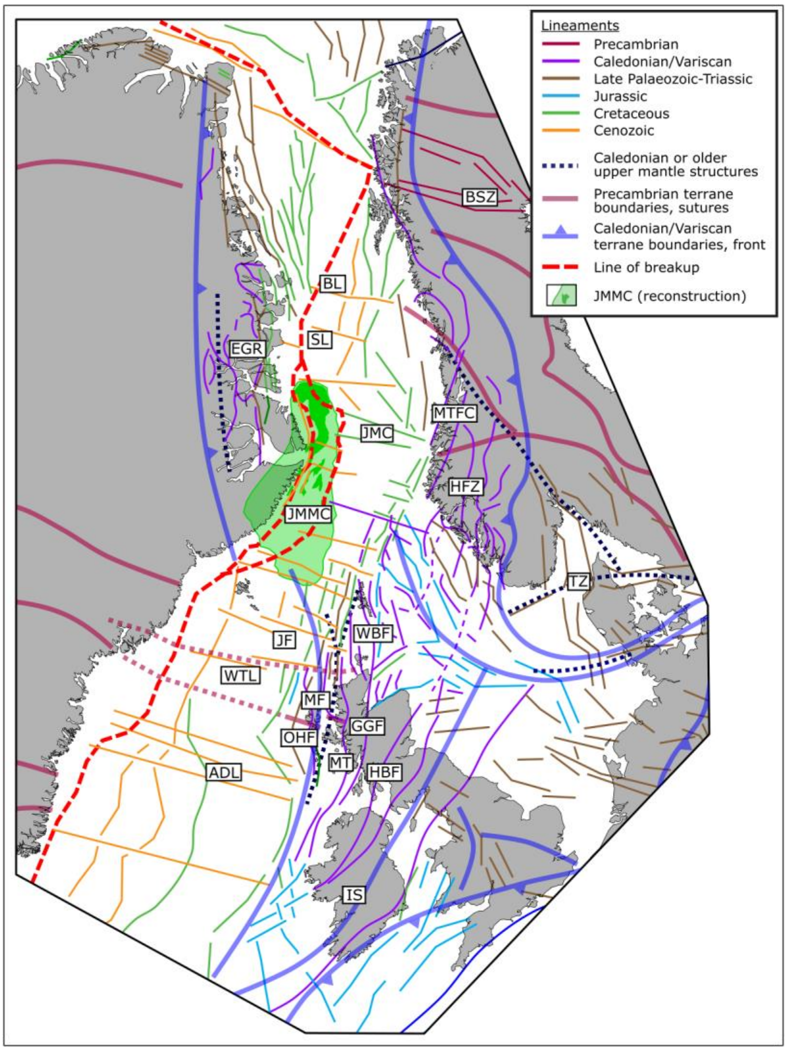

Figure 9: Basement terranes and lineaments of the NE Atlantic margins in a plate reconstruction at 60 Ma. Lineaments are coloured according to their main observed age of expression (Gernigon et al. this volume; Karson and Brooks 1999; Doré et al. 1999; Heeremans and Faleide 2004; Kimbell et al. 2005a; Guarnieri 2015; Fossen et al. 2017; Rotevatn et al. 2018; Holdsworth et al. 2019)***. Abbreviations: ADL, Anton Dohrn Lineament; BL, Bivrost Lineament; BSZ, Bothnian-Senja Shear Zone; EGR, East Greenland Rift System; GGF, Great Glen Fault; HBF, Highland Boundary Fault; HFZ, Hardangerfjord Fault Zone; IS, Iapetus Suture; JF, Judd Fault; JMMC, Jan Mayen Microplate Complex; JMC, Jan Mayen Corridor; MF, Mich Fault Zone; MTFZ, Møre-Trøndelag Fault Zone; MT, Moine Thrust; OHF, Outer Hebrides Fault Zone; SL, Surt Lineament; TZ, Tornquist Zone; WTL, Wyville-Thomson Lineament.***

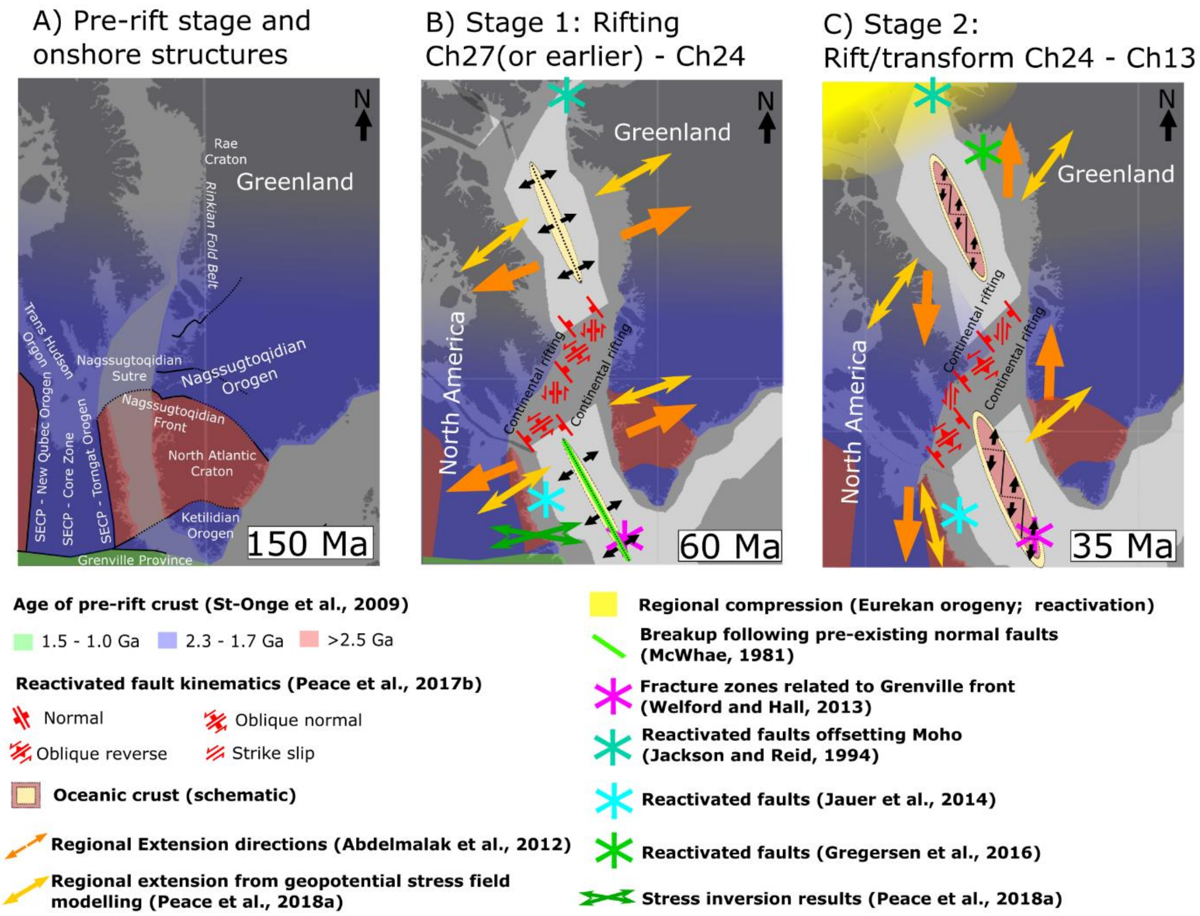

Figure 10: Schematic diagram summarising the kinematic and structural development, as well as seismic, mapping and modelling results of the Labrador Sea and Baffin Bay spreading system after (Peace et al. 2018b)***. A) Pre-rift configuration of North America and Greenland with graphical representations of the onshore structure and basement terrains of West Greenland (e.g. Wilson et al., 2006). B) Kinematic model for the first rift phase. C) Kinematic model for the second rift phase, after a change of stress field from ~SE-NW to N-S, at which the Ungava Transform Fault system develops as a result of the lateral offset between the Baffin Bay and Labrador Sea spreading centres.***

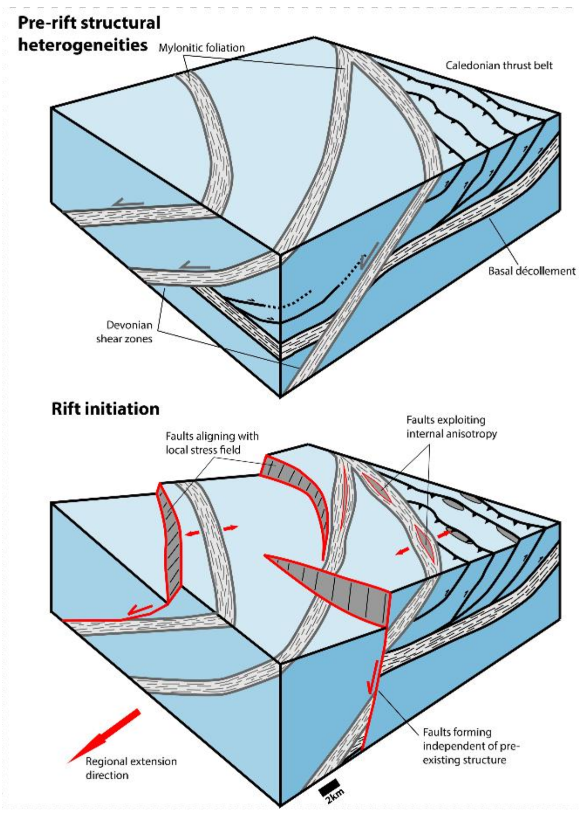

Figure 11: Schematic block diagram showing the Caledonian thrust belt and Devonian shear zones present within the lithosphere (above) and the various interactions with rift-related faults (below). After Phillips et al. (2016).

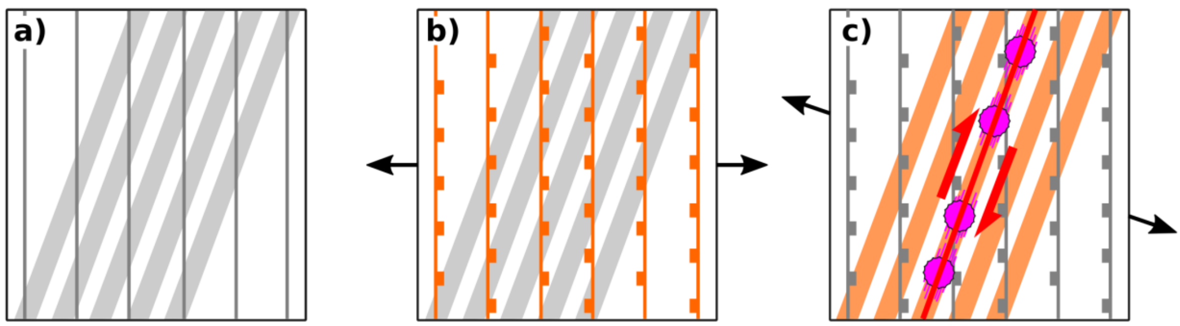

Figure 12: Simplified schematic diagram illustrating large-scale inheritance and reactivation in the NE Atlantic. The diagram is not to scale and is conceptual (i.e. does not show specific structures) (a) The initial shallow (crustal) Caledonian thrust fault systems (thin, dark grey lines) were oblique to the late Caledonian mantle shear fabric (light grey bands). (b)During early basin formation, the shallow Caledonian thrust faults were reactivated as normal faults (orange). Breakup, however, was not accommodated as the stress field was oblique to the stronger mantle shear fabric (light grey). (c) First later, after rotation of the stress field, the mantle shear fabric (orange) was favourably aligned allowing lithospheric breakup (red line). This may have been assisted by the formation of magmatic centres (magenta circles) and dykes exploiting lithospheric weaknesses

(Gernigon et al. this volume; Geoffroy et al. 2007) ***and/or strike slip motion, perforating the lithosphere prior to breakup*** **(Lundin and Doré 2018).**

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